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THE CEJA DEL RIO PUERCO: A BORDER FEATURE OF THE BASIN AND RANGE PROVINCE IN NEW MEXICO

II. GEOMORPHOLOGY

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ABSTRACT

The deformed beds of the Llano de Albuquerque are beveled and overlain by a cover of gravel, caliche, soil, and wind-blown sand 5–70 feet thick. Topographically the Llano de Albuquerque is a plain which slopes gently eastward toward the Rio Grande and is surrounded by escarpments and erosional slopes; it is a remnant of a once widespread erosion surface made up of coalescing pediments graded to the Rio Grande. At the time of formation of this surface, here named Ortiz, the Rio Grande had a grade about 500 feet above the present. On it were poured the great basaltic flows of the Mount Taylor region. The Ortiz surface has been dissected and largely destroyed by several successive lowerings of the grade of the Rio Grande. In the first period of dissection following Ortiz time the Rio Puerco gained its north-south course by piracy of certain western tributaries of the Rio Grande flowing in a southeasterly direction out of the Plateau Province. The vulcanism and erosion of later cycles are briefly summarized.

GENERAL STATEMENT

The discussion of the structure and stratigraphy of the Ceja del Rio Puerco in Part I of this paper[1] was definitely localized and limited to a consideration of the evidence found in two scarps—the

[1] *Jour. Geol.*, Vol. XLV (1937), pp. 801–28.

Ceja, which bounds the Llano de Albuquerque on the west, and part of the Rincones de Zia, which bound it on the north. The treatment wandered afield only to correlate the local Santa Fe beds with those of the type locality and to contrast the local pattern of faults with those of the immediately adjacent fault systems. However, any discussion of the geomorphology which hopes to be intelligible can hardly be so local. Its major emphasis must necessarily be on processes and events that are widespread and that affect both the Llano de Albuquerque and other areas throughout the length of the Rio Grande Depression from the San Luis Valley of Colorado southward to El Paso, Texas. These broad relationships are becoming clearer, but many of them await localized study. In this paper the main issues are (1) the nature of the surface of the Llano de Albuquerque, (2) the position of this surface in the sequence of geomorphic events, and (3) the origin of the Lower Rio Puerco by piracy.

THE RIO GRANDE DEPRESSION

The Rio Grande flows from the mountains of Colorado in a southerly course through southern Colorado and New Mexico in a series of open basins that, taken together, may be called the Rio Grande Depression. The northern part of this Depression lies within the Southern Rocky Mountain Province as usually defined[2] and the southern part within the Basin and Range Province, the boundary between the two provinces lying in central New Mexico. Recent work[3] in the area seems to indicate that this boundary is arbitrary. The Southern Rocky Mountains are divided into two prongs (Part I, *Jour. Geol.*, Vol. XLV [1937], Fig. 1). The eastern prong is the southern extension of the Sangre de Cristo Range, and the western prong consists of the Conejos and Jemez mountains with an intervening area of plateau. Between them the Rio Grande Depression extends as far north as Poncha Pass at the head of the San Luis Valley

[2] N. M. Fenneman *et al.*, "Physical Divisions of the United States," *U.S. Geol. Surv. Map* (1930).

[3] Summarized in Kirk Bryan, "Outline of the Geology and Ground-Water Conditions of the Rio Grande Depression in Colorado and New Mexico" (mimeographed report, Rio Grande Joint Investigation, Water Resources Committee, National Resources Committee, 1936–37).

and carries the structure and topography of the Basin and Range Province far into the mountain region.

In Colorado the trough of the Rio Grande is called the San Luis Valley and is usually and quite arbitrarily classified as one of the "parks." In New Mexico the Depression includes the basalt plateau west of Taos, the Española Valley, the broad plains west of Santa Fe, the basalt and andesite plateau known as the Cerros del Rio or Pankey's Mesa, and southward a series of broad basins bordered east and west by sharply serrate ranges. The Albuquerque-Belen basin in central New Mexico is one of this series. It is approximately 20 miles wide and 80 miles long and differs from the rest in that along its western margin the Basin and Range Province borders on the Plateau Province with only moderate differences in elevation. Elsewhere the Plateau fronts on the Basin and Range Province in dominating cliffs, as in the Mogollon Escarpment, the Grand Wash Cliffs, or the western margin of the High Plateaus of Utah. The area here considered is the northwestern corner of the Albuquerque-Belen Basin.

THE LLANO DE ALBUQUERQUE

Between the Rio Puerco on the west and the Rio Grande on the east is a long, narrow, pointed plain, the Llano de Albuquerque, approximately 70 miles long and 12 miles wide at the northern end, where its width reaches a maximum. The smooth, grassy surface of the plain slopes gradually southward and eastward, dropping from an elevation of 6,750 feet at the north to about 5,000 feet at its southern end. This plain is bounded on the north by the Rincones de Zia, an area of badlands lying just south of Jemez Creek. On the south, it narrows to a tongue of high ground between the confluent valleys of the Rio Grande and the Rio Puerco. The western edge is the escarpment known as the Ceja del Rio Puerco or "Eyebrow of the Dirty River" which breaks off sharply to the dissected valley of the Rio Puerco. Eastward, the plain is extensive and the slope is gentle until it breaks at a lower, less regular, and less prominent scarp called the Cejita Blanca or "Little White Eyebrow." Below this minor scarp lies a lower plain that merges with sandy wastes and dissected terraces leading down to the valley of the Rio Grande

on the east. Southward, the Llano de Albuquerque loses its eastward slope and, at its southern end, becomes a flat-topped, low plateau.

The northern 30 miles of the scarp which bounds the Llano de Albuquerque on the west is coextensive with a system of en echelon faults having their downthrow to the east.[4] Such faulting should produce an irregular, east-facing scarp. The present scarp faces west. It, then, must be a fault-line scarp and clearly a reverse fault-line scarp, since it faces in a direction opposite to that of the original movement. Such a border between the Basin and Range and Plateau provinces is unique. The southern 40 miles of the scarp is a purely erosional feature separated from the fault system and the Plateau Province by the broad valley of the Rio Puerco.

Genetically, the Llano de Albuquerque may be considered from two viewpoints. Its surface is generally coincident with the bedding of a cover of caliche, fine-grained wind-blown soil, and a sheet of gravel. From this point of view it is a depositional surface and on this account has heretofore been regarded as the remnant of the depositional surface[5] of a great alluvial fill in the Rio Grande Basin. On the other hand, the gentle, eastward-sloping surface cuts the underlying, deformed Santa Fe beds, and, below the thin mantle of cover, there is an angular unconformity. In the present discussion this erosional surface is regarded as of primary importance.

In addition to the alluvium which covers the erosional surface of the Llano de Albuquerque, there is a considerable quantity of basalt at two localities. The small field of basalt derived from the five Albuquerque volcanoes lies mostly on the surface of the Llano, but it also extends down onto the next lower erosion surface, known as the Segundo Alto or Segundo Banco. It must, therefore, have been erupted after the dissection of the Llano had begun. West of Isleta and Los Lunas, hills composed in part of basalt rise above the surface of the Llano. These hills have been carved from basalt flows

[4] See Part I of this paper, *loc. cit.*

[5] Bryan, "Geology of the Vicinity of Albuquerque," *loc. cit.*, pp. 5–24; W. T. Lee, "Water Resources of the Rio Grande Valley in New Mexico," *U.S. Geol. Surv. Water Supply Paper 188* (1907), pp. 20–24, and Fig. 1; N. L. Darton, "Guidebook of the Western United States, Part C: The Santa Fe Route," *U.S. Geol Surv. Bull. 613* (1915), Fig. 16.

interbedded with the Santa Fe and are residuals above the surface of the Llano. However, in this vicinity there is an area of much more recent volcanic activity. Northwest of Los Lunas there are cinder cones and a basalt flow, whose fresh aspect and slightly eroded surface indicate that they are no older and are, perhaps, younger than the basalt of the Albuquerque volcanoes. A tongue of this younger basalt extends eastward down a shallow valley cut in the surface of the Llano north of the state highway. This flow consequently must have been erupted after, and perhaps long after, the dissection of the Llano had begun.

That the cover, including alluvium and basalt, rests unconformably on the Santa Fe in this area can be clearly demonstrated, for the beds of the mantle slope gently with the surface whose maximum gradient does not exceed 3°, whereas the Santa Fe beds may dip as much as 10°. In other localities, however, the Santa Fe beds are so nearly horizontal that the discordance cannot be easily detected. The lithology and texture of the gravel cover differentiate it from the Santa Fe beneath. On the whole, the three members of the Santa Fe are much finer grained than the gravel cover. Only a few of the gravels of the Upper Buff Member have as large pebbles as those found in the cover. Many pebbles are andesite, whereas the gravel of the cover contains a large percentage of quartz, quartzite, and petrified wood.

The conclusion that a great erosion surface bevels the Santa Fe beds in this area is further reinforced by the presence of caliche which is from 2–20 feet thick. This is the "marl" of Herrick's[6] paper and can usually be divided into a lower concretionary portion, a middle, massive, finely granular mass, 50–60 per cent calcium carbonate, and an upper platy portion, a few inches to 2 feet thick. As Breazeale[7] has shown, caliche may be formed in more than one way. Disregarding detailed processes, it must be conceded that thick beds are formed on surfaces of low relief that are stable over a long period of time. Such conditions are best attained on erosion surfaces

[6] C. L. Herrick and D. W. Johnson, "The Geology of the Albuquerque Sheet," *Univ. of New Mexico Bull.*, Vol. II, Part I (1909), p. 4. See also *Denison Univ. Scientific Laboratories Bull.*, Vol. XI.

[7] J. F. Breazeale and H. V. Smith, "Caliche in Arizona," *University of Arizona, Agric. Exper. Stat. Bull. 131* (1930), pp. 419–41.

over which streams play at will. Caliche as an indicator of mature erosion surfaces has long been used by the first-named author of this paper, whose views are similar to those recently expressed by Price.[8]

The gravel of the cover is subangular and rudely sorted, indicating deposition by ephemeral and intermittent streams. That these streams originated to the west and northwest is obvious from the abundance of pebbles of quartz, quartzite, jasper, and petrified wood and the nearly complete absence of andesite pebbles that might have come from the north.

Conglomeratic zones in the Abo sandstone and Chupadera formation of the Nacimiento Uplift may have furnished part of this quartzitic material, but much is probably derived from the coarser zones of the Mesozoic beds of the San Juan Basin. The character and appearance of the fragments of petrified wood indicate that they are quite certainly derived from Cretaceous and early Tertiary beds of the Plateau area to the west.

The western margin of the Llano, on the brink of the Ceja, from the vicinity of Sand Hill and the head of the Canyada Moquino southward, is marked by the presence of a line of sand hills 50–70 feet high. These hills are typical cliff dunes, and the wind-blown sand which forms them thins rapidly eastward, so that they disappear as a topographic feature within a few hundred feet. In this locality the loosely consolidated sands of the Upper Buff Member of the Santa Fe furnish large quantities of fine material to the rills and intermittent streams below the scarp of the Ceja, from whose dry channels the prevailing westerly winds pick it up, sweeping it eastward and upward through the gulches of the badlands onto the edge of the plain. Despite their relative freedom from vegetation, the dunes migrate eastward very slowly, for the upward draft of air stops at the edge of the plain, and the dunes are being continually undermined and their material carried westward again by rainwash down the same gullies from which it was first lifted by the wind.

The surface of the Llano is dissected by a few broad, shallow valleys which trend southeast (Fig. 1). There is a progressively

[8] W. Armstrong Price, "Reynosa Problem of South Texas and the Origin of Caliche," *Bull. Amer. Assoc. Petrol. Geol.*, Vol. XVII (1933), pp. 488–522, and discussion in *ibid.*, pp. 1277–81.

poorer development of the valleys toward the south and at the southern end of the plain they give place to slight hollows and depressions. This change is due partly to the decreasing width of the Llano, partly to the small basalt field of the Albuquerque volcanoes and that near Los Lunas, and partly to unknown causes. These valleys have been cut by streams never completely graded to the Rio Grande drainage. A heavy, prolonged rain may cause the larger of them to run for several hours, and occasionally water reaches the Rio Grande. The surface soil, however, is so porous that ordinarily there is little runoff save where the caliche is close to the surface.

FIG. 1.—Broad, shallow valleys trending southeastward and, as seen in the middle of the view, beheaded by the retreat of the Ceja del Rio Puerco.

The gradient of these southeasterly flowing, intermittent streams on the Llano, tributary to the Rio Grande, is much less than that of those intermittent streams, tributary to the Rio Puerco, which flow west down the Ceja slope. The inevitable consequence of this difference in gradient is the slow eastward retreat of the Ceja, the divide which separates them.

THE ORTIZ SURFACE

The Rio Grande Depression is characterized by the presence of remnants of once broad erosion surfaces. These remnants stand at widely different elevations, and their degree of preservation is variable from place to place. However, they all slope toward the general position of the Rio Grande and belong to a succession of erosional surfaces representing successive stabilized grades of this master-

stream. In the region south of White Rock Canyon the oldest and highest of these remnants appear to be parts of the Ortiz surface, a widely developed and excellently preserved surface in the vicinity of the Ortiz Mountains, first described by Ogilvie.[9] Although Ogilvie described but one surface in the type area, there appear, in fact, to be three, and it is the highest of these which is here named the Ortiz surface. It slopes outward on all sides away from the mountain and, if restored to its original condition, would be a low cone, the "conoplain" of Ogilvie. It cuts the hard rocks at the foot of the mountains and extends westward across the deformed Santa Fe. In places this surface is little dissected, and its gradient, projected toward the Rio Grande in the type area, reaches the river about 500 feet above its present grade.[10]

North of the White Rock Canyon, in the Española–Santa Fe Basin and in the bordering re-entrants, there are many remnants of erosion surfaces apparently graded to the Rio Grande at elevations of 1,000–2,000 feet above its present grade. Whether these surfaces or any one of them is the equivalent of the Ortiz remains to be determined.[11] Apparently deformation or vulcanism or both have given this northern area a more complicated erosional history than the area south of White Rock Canyon.

In the Albuquerque-Belen Basin, the Ortiz surface is represented by the Alto de Mesa Santa Ana and by the Llano de Albuquerque. On the Mesa Santa Ana the surface is buried beneath a lava capping and the topography is not well known. The more accessible Llano is shown on the reconnaissance map of the Albuquerque Quadrangle, surveyed by the United States Geological Survey in 1888. Its surface slopes southeastward toward the Rio Grande (Fig. 2). The gradients of the surface decrease toward the south, varying from about 85 feet per mile at the northern end of the Llano (Profile *I*,

[9] Ida H. Ogilvie, "The High-Altitude Conoplain: A Topographic Form Illustrated in the Ortiz Mountains," *Amer. Geol.*, Vol. XXXVI (1905), pp. 27–34; see also Douglas Johnson, "Rock Fans of Arid Regions," *Amer. Jour. Sci.*, Ser. 5, Vol. XXIII (1932), p. 394.

[10] To be described by Bryan and Upson in a forthcoming paper, "Geomorphology of the Santo Domingo Valley, New Mexico."

[11] E. C. Cabot, "Fault Border of the Sangre de Cristo Mountains North of Santa Fe, N.M.," *Jour. Geol.*, XLVI (1938), 88–105.

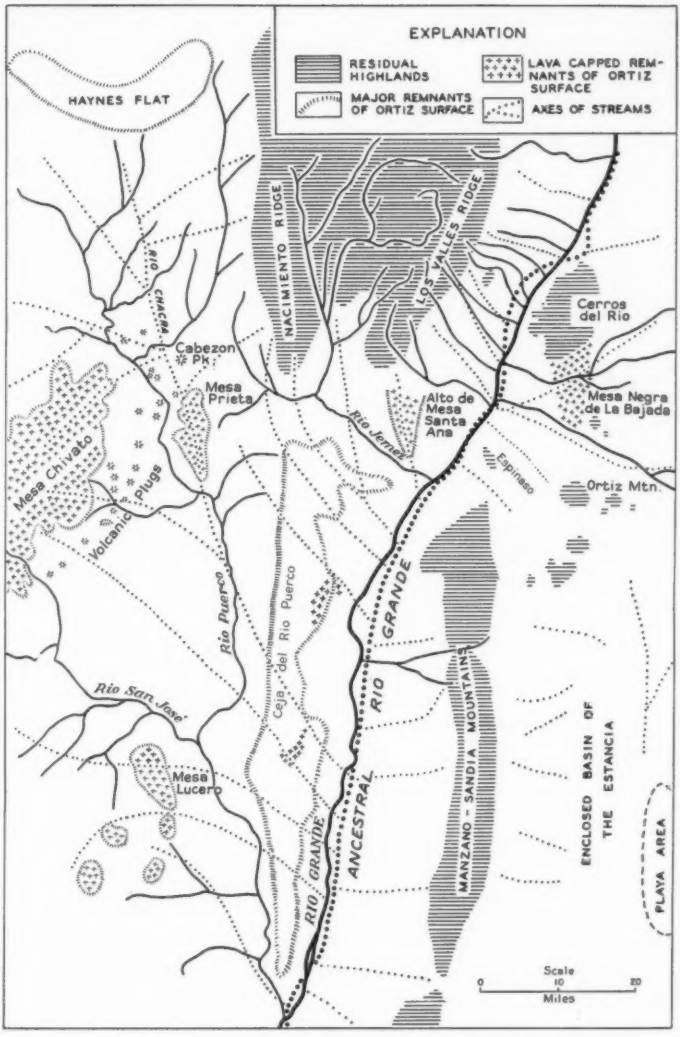

FIG. 2.—Map of central New Mexico in Ortiz time showing the ancestral Rio Grande, the hypothetical Rio Chacra, and other hypothetical stream courses, and their relation to present drainage.

Fig. 3) to about 60 feet per mile in the vicinity of Albuquerque (Profile *III*, Fig. 3). These gradients, as determined from the contours of the Albuquerque sheet, may be in error, but they are such as can be reasonably expected. The gravel of the cover is angular and unsorted and of the type that would be laid down by ephemeral streams having such grades. In projecting these gradients southeastward to the river they have been slightly flattened. Because of

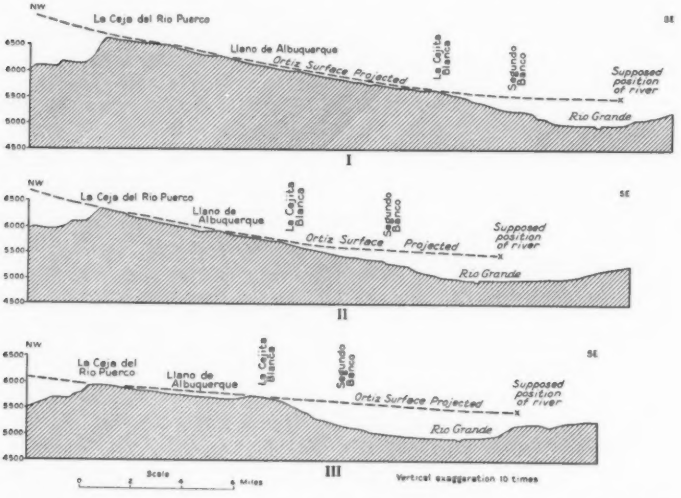

FIG. 3.—Profiles across the Llano de Albuquerque from north to south, at intervals of approximately 10 miles, showing the grades of the Ortiz surface and the presumed positions of the Rio Grande of Ortiz time.

the progressive decrease in gradient toward the south, there is a corresponding progressive increase in the altitude of the projected grade above the present Rio Grande which increases from about 500 feet at the north end of the Llano (Profile *I*, Fig. 3) to about 550 feet in the vicinity of Albuquerque (Profile *III*, Fig. 3). Considered ideally, the altitude of the projected surface above the present Rio Grande should be uniform. Such variation from the ideal as has just been described may possibly be due to errors in the profiles or to irregularities in the Ortiz surface, but it seems more likely that there

has been a westward shift in the position of the Rio Grande and that in Ortiz time the river lay near the positions marked *X* in the profiles.

The great basalt sheets of the Mount Taylor area slope outward from the mountain in all directions. They rest on the eroded edges of the underlying deformed Cretaceous rocks, as described by many authors and demonstrated graphically in the recently published structural map by Hunt[12] and by others. Before the lavas were extruded, this erosion surface must have been connected with other remnants which together formed a broad surface of major physiographic importance. As shown in Figure 4, this surface has a gradient such that it may easily be the equivalent of the Ortiz surface. In the Mesa Prieta, the Mesa Chivato, and the Alto de Mesa Santa Ana, the Ortiz surface is covered by basaltic lava. Part of this lava may have been poured out before the surface was wholly complete, preventing erosion and preserving the part covered at elevations somewhat above normal. On the other hand, some of the lava may have been erupted after dissection of the surface had begun. This is certainly the case with the lava of the Albuquerque volcanoes, which extends down onto the Segundo Alto.

East of the Rio Grande in this area the Ortiz surface is absent except for remnants of surfaces so far not certainly identified.

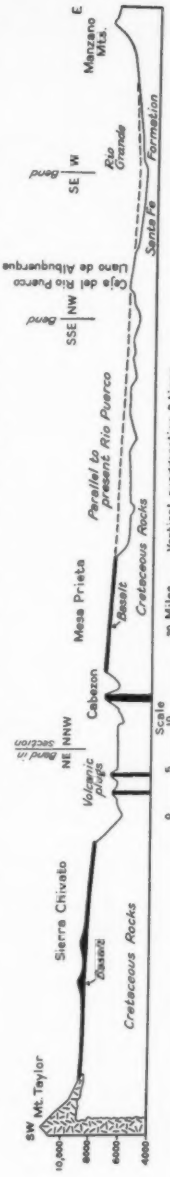

FIG. 4.—Profile showing gradient of Ortiz surface from Mount Taylor northeast to Cabezon, thence down the general gradient of the Rio Chacra to the Rio Grande, and up the Ortiz surface to the Manzano Range.

[12] Charles B. Hunt, "Geology and Fuel Resources of the Southern Part of the San Juan Basin, New Mexico," *U.S. Geol. Surv. Bull. 860-B* (1936), pp. 1–80, and maps.

The short streams flowing out of the Sandia and Manzano mountains are powerful in flood and must have developed an erosion surface which, in fact, is well preserved at the north end of the Sandia Range. Under changed conditions (see above) they would be equally capable of destroying most of the surface and reducing it to the gradient of the next lower surface. This seems to be the most reasonable explanation for the general absence of the Ortiz surface east of the Rio Grande and the consequent asymmetry of the valley brought out in Figures 4 and 5. In 1909 Bryan[13] noted this asymmetry and, believing that the highest surfaces on either side of the Rio Grande were equivalent, postulated differential uplift of the western plains with respect to the eastern. There is no other evidence of such an uplift. In fact, the downthrow on the fault system along the Ceja is to the east, showing that the area west of the Rio Grande has been carried downward instead of upward. Nor is the evidence of equivalence of surfaces compelling. The eastern plains, the Llano Sandia, have on their highest surfaces a capping of caliche about $2\frac{1}{2}$ feet thick. This is in marked contrast to the caliche on the Llano de Albuquerque, which is from 2–20 feet thick. In the paper of 1909[14] Bryan explained this discrepancy in thickness on the ground of a higher percentage of limestone pebbles in the gravels below the caliche of the Llano de Albuquerque. Further field work indicates that there is no such preponderance of limestone pebbles or other calcium carbonate rock. The thickness of the caliche appears to be roughly a measure of the completeness of the erosion surface. The older and more complete Ortiz surface has the thicker plate of caliche.

THE PIRACY OF THE LOWER RIO PUERCO

The Rio Puerco, the major western tributary of the Rio Grande, flows nearly due south 140 miles from the town of Cuba to the town of La Joya, where it joins the Rio Grande. En route it traverses a variety of formations. It flows out of the granite of San Pedro Mountain, and passes quickly across the upturned edges of the Mesozoic and late Paleozoic sedimentary rocks onto the nearly flat-

[13] "Geology of the Vicinity of Albuquerque," *loc. cit.*, pp. 11–12 and Fig. 6.

[14] *Ibid.*, pp. 19–23.

lying Wasatch and Nacimiento beds, then onto the Lewis, Mesa Verde, and Mancos of Cretaceous age. It continues to cut its valley generally southward in the Mancos past Cabezon and the Mesa Prieta, works again for some distance on the Nacimiento beds, and flows the remaining distance across the Santa Fe formation before emptying into the Rio Grande.

The Rio Puerco possesses two important tributaries, both entering it from the west. The Rio San José empties into the master-stream 41 miles above its mouth and the Arroyo Chico about 103 miles above the same point. The juncture of the Arroyo Chico and the Rio Puerco has been selected as marking the division between the "upper" and "lower" Rio Puerco because it seems probable that an Upper Rio Puerco and Lower Rio Puerco once actually existed as separate streams possessing a length and position comparable to those of these two portions of the present stream.

At the end of Ortiz time and before the development of the La Jara surface[15] the western part of the Albuquerque-Belen Basin was drained by a series of southeasterly flowing streams, among them a hypothetical master-stream called the Rio Chacra (see Fig. 2), which flowed southeastward from the San Juan Basin between the residuals of Mount Taylor and those of the Nacimiento Uplift. The Rio Chacra probably headed in the vicinity of Haynes Flat and, its course directed by these residuals, flowed across the Mesa Prieta. The present southerly course of the Rio Puerco represents a change in direction of drainage, due probably to two successive piracies by which the Rio Chacra was converted into the Rio Puerco. The piracy by which the Upper Rio Puerco came into existence[16] resulted from the headward capture of upper tributaries of the Rio Chacra by a lower tributary working headward first in a belt of weak Mancos shale and afterward in the Lewis shale. The mechanism by which the Lower Rio Puerco came into existence must have been much the same, with certain minor differences. The capturing, in this instance, was probably done by a northerly tributary to a westerly tributary of the ancestral Rio Grande. This westerly tribu-

[15] Kirk Bryan and Franklin T. McCann, "Successive Pediments and Terraces of the Upper Rio Puerco in New Mexico," *Jour. Geol.*, Vol. XLIV (1936), pp. 160–64.

[16] *Ibid.*, pp. 154–58.

tary flowed eastward and entered the river at or near the mouth of the present Rio Puerco. Its northerly tributary must have worked headward in the weak Santa Fe beds until it captured the main channel of the Rio Chacra, which probably flowed across the Llano de Albuquerque in the vicinity of Los Lunas. Thus the lower part of the Lower Rio Puerco was fixed in the Santa Fe beds east of the zone of faulting that bounds the Albuquerque-Belen Basin. The reasons for the inception and location of the capturing tributary at this point are not clear. One would expect that the fault zone would be the zone of weakness in the area. North of Los Lunas the course of the Rio Chacra as far north as Cabezon became the course of the upper portion of the Lower Rio Puerco, and the present course of the stream is the incised course of the Rio Chacra with a slight shift to the west.

LATER EROSION SURFACES

Subsequent to the cutting of the Ortiz surface, the Rio Grande successively lowered its grade and remained stabilized at each successive grade for considerable periods of time. These new grades may have been caused by a series of uplifts that affected the greater part of New Mexico, or they may have been controlled in part by the position of hard-rock sills in the lower course of the river in Texas, or by the position of sea-level. The recent work of Price[17] demonstrates that the delta of the lower Rio Grande was the western border of an active geosyncline in Pliocene and Pleistocene time, and movement in this area may have affected the Rio Grande far upstream.

With each successive lowering of grade the tributary streams were also incised. When the grade of the Rio Grande was stabilized, that of each of its tributaries was also stabilized, and pediments or broad plains of erosion were cut. No one of these was so well developed or so extensive as the Ortiz surface. Two of these later surfaces are preserved in the Ceja area. The higher of them is locally called the Segundo Alto or Banco and the lower the Primer Alto or Banco. Although correlation is not entirely satisfactory, these probably correspond to the La Bajada and Peña Blanca pediments of the Santo

[17] Price, *op. cit.*, pp. 488–522.

Domingo area,[18] the Tio Bartolo and Valle la Parida pediments of the Socorro area,[19] and the La Jara and Rito Leche pediments of the Cuba area.[20]

West of the escarpment of the Ceja, there are extensive remnants of pediments between the deep gulches or canyadas of the main streams tributary to the Rio Puerco. As shown in Figure 5, these remnants are sloping plains mantled with alluvium and wind-blown sand. They slope toward the Rio Puerco, but their character is lost in the intricate topography of the mesas and ridges near that stream. The largest area is known as Pantaleon and is shown in Figure 4 (Part I of this article, *loc. cit.*). It seems likely that these surfaces

FIG. 5.—Remnant of pediment just north of the Canyada Navajo; the escarpment of the Ceja del Rio Puerco in the background.

are graded to a level of the Rio Puerco about 125 feet above its present grade. However, there are other surfaces along this stream which complicate the picture and make reconnaissance correlation hazardous.

LATER TERRACES

Terraces at lower levels than the pediments occur at various localities along the Rio Grande and its principal tributaries. These terraces are of two kinds: (1) those due to alternate deposition of alluvial fans by tributaries and lateral planation by the river and

[18] Bryan and Upson, unpublished data.

[19] Bryan, "Pediments Developed in Basins with Through Drainage as Illustrated by the Socorro Area, New Mexico," *Abst. Geol. Soc. Amer. Bull.*, Vol. XLIII (1932), pp. 128–29.

[20] Bryan and McCann, *op. cit.*, pp. 145–72.

(2) those formed during previous stabilizations of the main river at grades higher than the present. Terraces of the first type have no chronologic significance. Of the latter type, terraces from 10 to 100 feet above the present river grade are known and doubtless represent the events of late Pleistocene time.

ACKNOWLEDGMENTS.—The field work was financed in part by a grant from the Milton Fund of Harvard University. Messrs. George R. Crowl and Douglas Scott, as volunteer assistants, carried on the plane-table work, leaving Bryan and McCann much free time for general reconnaissance and the stratigraphic and structural work. The month of August, 1933, was spent in two camps on the brink of the Ceja. Acknowledgment is due to Mr. Hugh M. Bryan, who loaned part of the equipment, and to Mr. Field Thompson, who allowed the party to procure water from the headquarters well of the Thompson Ranch.

A STUDY IN ROCK-WEATHERING

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ABSTRACT

Four districts furnished samples for a study in rock-weathering. Mineralogical data and fourteen new chemical analyses are presented in tables and diagrams. The granite gneiss of Morton, Minnesota, shows a marked loss of soda and lime in early stages of weathering but a more gradual loss of potash and baria. The progress of the decomposition is interpreted from a series of six samples of residual clay. Kaolinite is an end product.

New chemical analyses of material from the Medford, Massachusetts, diabase are presented as a check on the older data. Ferrous and ferric iron determinations on samples of diabase and of glacial till suggest that the till is less oxidized than the underlying diabase, indicating preglacial weathering of the diabase.

A diabase on the north shore of Lake Superior, Minnesota, is weathered locally to a depth of 40 feet but shows little chemical change except oxidation.

An amphibolite from the Black Hills, South Dakota, yielded beidellite or related clay minerals by decomposition of its hornblende. Calcite has been leached, and other minerals attacked.

On the basis of the present and earlier studies, a *mineral-stability series in weathering* is proposed. The arrangement of the common rock-forming minerals in this series is found to coincide with the reaction series. Experimental work on the attack of silicate minerals by water is in accord with the suggested mineral-stability series.

INTRODUCTION

The effects of rock-weathering have long been observed and studied, and the interest in this field of geological investigation is indicated by a voluminous literature. With the increasing emphasis on sedimentation and on the origin and development of soils, the basic contributions of studies in weathering assume even greater significance.

The present study is an attempt to analyze the chemical and mineralogical changes produced in the weathering of certain rocks from a few localities selected chiefly because of their accessibility to the writer or because suitable samples were available. The rocks and their weathering products studied are the following: (1) granite gneiss from the vicinity of Morton, Minnesota; (2) diabase from Medford, Massachusetts; (3) diabase from the north shore of Lake Superior, Minnesota; (4) amphibolite from the Black Hills, South Dakota.

The writer gratefully acknowledges his indebtedness to Professors

F. F. Grout and G. A. Thiel, of the University of Minnesota, and to Dr. R. B. Ellestad, of the Rock Analysis Laboratory of that institution for many suggestions and criticisms which aided in the development of the problem. Mr. Robert Grogan assisted in the field work. Thanks are due to Professor C. L. Baker and Dr. F. E. Turner, of the Agricultural and Mechanical College of Texas, for their critical reading of the manuscript.

METHODS OF STUDY

CHEMICAL ANALYSIS

The new chemical analyses presented in this paper were made in the Rock Analysis Laboratory in the Department of Geology of the University of Minnesota. The methods routinely employed in that laboratory are essentially those described by Hillebrand and Lundell.[1]

SPECIFIC GRAVITY

Determinations of specific gravity were made with the fused silica pycnometer designed and described by Ellsworth.[2] Tables and charts prepared by Dr. R. B. Ellestad facilitated the calculations and the application of buoyancy corrections.

MINERALOGICAL STUDY

Most students of weathering have relied on chemical data and on calculations based on these data for a quantitative measure of rock decomposition. Inferences can be made from chemical analyses as to the mineralogical changes which have taken place. Nevertheless, it seems desirable to attempt some correlation between the changes indicated by the analytical results and the actual mineral changes. In the mineralogical study a combination of methods was used. A technique involving the use of a heavy liquid and a multiple-type of electromagnetic separator may warrant brief description. The general procedure is used extensively in sedimentation and in accessory mineral studies.

Samples of both fresh and weathered material were crushed to

[1] W. F. Hillebrand and G. E. F. Lundell, *Applied Inorganic Analysis* (New York: John Wiley & Sons, 1929).

[2] H. V. Ellsworth, "A Simple and Accurate Constant-Volume Pyknometer for Specific Gravity Determinations," *Min. Mag.*, Vol. XXI (1928), pp. 431–36.

reduce the rock aggregate as much as possible to monomineralic particles. The pulverized samples were washed with water, and the fraction remaining in suspension after a settling period of three minutes in a 5-inch column of water was removed by siphoning.

The washed samples after being dried were separated by means of bromoform (sp. gr. = 2.88) into a light and heavy fraction. The light fractions were bottled and reserved for study. The heavy-mineral fractions were further split by electromagnets into four separates of different magnetic susceptibility. These separates were weighed, and a portion of each, obtained with a micro-splitter, was mounted in Canada balsam. The relative proportions of the minerals in each separate were determined by counting. From the frequency percentages, the approximate weight of each mineral in all four separates was obtained and converted into the final percentage representing the approximate percentage by weight of the mineral in the heavy-mineral fraction.

The heavy-mineral technique was introduced as an experiment. It affords the advantage of permitting both fresh and weathered material to be treated by a single procedure.

CALCULATIONS AND DIAGRAMS

A variety of diagrams have been used by geologists to show the effects of weathering. The "straight-line" diagram[3] is used to show the relative gains and losses during rock alteration when no single constituent is assumed to have remained constant. The *Al-C-Alk* relationship proposed by Osann[4] is based on the proportions of alumina, lime, and alkalies. The molecular percentages of Al_2O_3, $(Ca,Ba)O$, and $(Na,K)_2O$ are recalculated to 30 and are plotted on a triangular diagram. According to Osann, all igneous rocks should fall to the right of the line which divides the triangle (Fig. 8) into two fields. Sedimentary and parametamorphic rocks fall in the field to the left of the line. Other calculations and diagrams used for single groups of analyses are discussed in appropriate places.

[3] C. K. Leith and W. J. Mead, *Metamorphic Geology* (New York: Henry Holt & Co., 1915), p. 288.

[4] Rosenbusch-Osann, *Elemente der Gesteinslehre* (4th ed.; Stuttgart, 1923), pp. 96–104, Pl. III.

MORTON GNEISS

GENERAL STATEMENT

Residual clays in the vicinities of Morton and Redwood Falls, Minnesota, were formed by pre-Cretaceous weathering of a granite gneiss. The fresh gneiss is well exposed in a number of quarries which have been described recently by Thiel and Dutton.[5] The residual clays are overlain by Cretaceous sediments or by glacial drift in places where the Cretaceous has been eroded. The clays of this region have been studied by Winchell,[6] by Upham,[7] by Hall,[8] by Meinzer,[9] and more recently by Grout.[10] Six samples of residual clay were studied for comparison with the fresh Morton gneiss. The locations of the samples are shown in Figure 1, and chemical analyses are given in Table 1. For comparison, analyses of fresh and decomposed gneiss previously published are included in the table.

FRESH MORTON GNEISS

Location and description of sample.—The Morton quarry of the Cold Spring Granite Company (Sec. 31, T. 113 N., R. 34 W.) is the largest in the district and affords the best possibilities for sampling. The gneiss is pink to dark gray. Quartz, pink to red potash feldspar, light-colored plagioclase, biotite, and magnetite are easily recognizable. The dark- and light-colored minerals are more or less segregated into bands. The texture is porphyritic and in some places pegmatitic. Pegmatites, quartz veins, and fine-grained granitic dikes of reddish color cut the gneiss.

Fourteen pounds of chips were taken from blocks and from quarry ledges and walls. The final sample (No. 1) was a composite of mate-

[5] G. A. Thiel and C. E. Dutton, "The Architectural, Structural and Monumental Stones of Minnesota," *Minn. Geol. Surv. Bull. 25* (1935), pp. 88–95.

[6] N. H. Winchell, *Geol. and Nat. Hist. Surv. Minn., 2d Ann. Rept.* (1874), p. 163; and *Geol. Minn.*, Vol. I (1882), p. 571.

[7] Warren Upham, "The Geology of Renville County," *Geol. and Nat. Hist. Surv. Minn.*, Vol. II (1888), pp. 190–204.

[8] C. W. Hall, "The Gneisses, Gabbro-Schists, and Associated Rocks of Southwestern Minnesota," *U.S. Geol. Surv. Bull. 157* (1899).

[9] C. W. Hall, O. E. Meinzer, and M. L. Fuller, "Geology and Underground Waters of Southern Minnesota," *U.S. Geol. Surv. Water-Supply Paper 256* (1911), pp. 304–24.

[10] F. F. Grout, "Clay and Shales of Minnesota," *U.S. Geol. Surv. Bull. 678* (1919), pp. 78–80, 216–22.

rial collected on different days by three men and is considered to be fairly representative of the gneiss in the quarry. Whether or not the sample is representative of all the bedrock in this portion of the

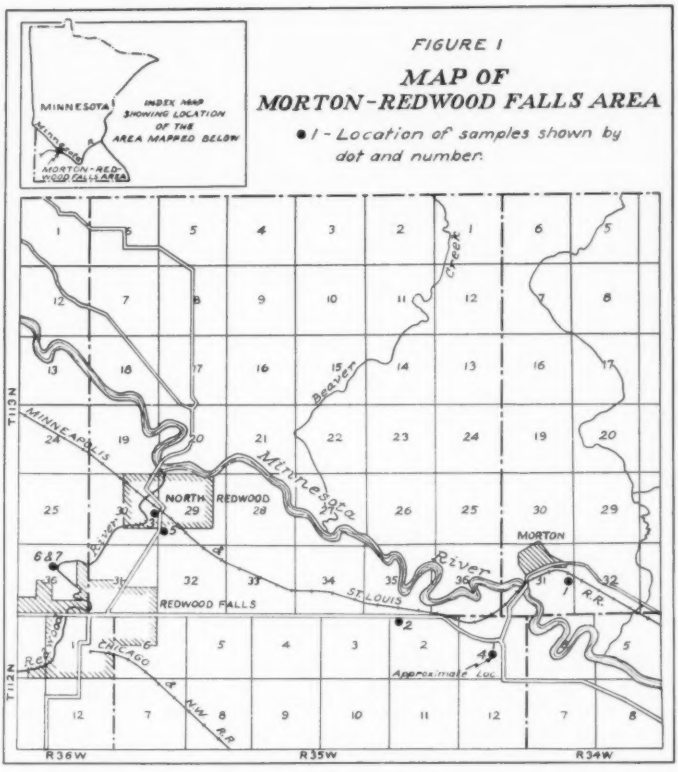

FIG. 1.—Map of Morton–Redwood Falls area

valley is open to some doubt. The selection of material would influence the results. However, the origin of the gneiss and the field relationships suggest that some differences may be expected in the parent-rock from which the residual clays in the Morton–Redwood Falls area have been derived.

The Morton gneiss is an injection gneiss—a hybrid formed by the invasion of schists (?) by a granitic magma. In smaller quarries about 1½ miles north of North Redwood (NW. ¼ Sec. 29, T. 113 N., R. 35 W.) banded gneiss similar in appearance to the rock in the Morton quarries grades into medium-grained granite.

TABLE 1

CHEMICAL ANALYSES OF FRESH AND WEATHERED GNEISS FROM MORTON–REDWOOD FALLS AREA

	1	2	3	4	5	6	7	8	9
SiO_2	71.54	69.89	68.09	61.75	70.30	57.53	55.07	63.61	60.61
Al_2O_3	14.62	16.54	17.31	18.58	18.34	23.57	26.14	16.71	18.12
Fe_2O_3	0.69	2.33	3.86	1.69	1.55	3.05	3.72	5.69	7.51
FeO	1.64	0.34	0.36	4.11	0.22	3.59	2.53	2.78	
MgO	0.77	0.30	0.46	0.76	0.21	0.41	0.33	1.63	1.14
CaO	2.08	0.06	0.06	0.16	0.10	0.05	0.16	4.03	0.03
Na_2O	3.84	0.43	0.12	0.10	0.09	0.06	0.05	1.68	0.54
K_2O	3.92	5.34	3.48	3.54	2.47	0.35	0.14	2.49	3.56
H_2O+	0.30	4.00	5.14	5.56	5.58	8.82	9.75	0.61	7.20*
H_2O-	0.02	0.35	0.47	0.35	0.30	0.70	0.64	.....	0.28
CO_2	0.14	0.21	0.05	1.84	0.20	0.77	0.36	.....	.....
TiO_2	0.26	0.14	0.34	0.92	0.21	0.87	1.03	.....	1.30
P_2O_5	0.10	0.04	0.03	0.09	0.04	0.08	0.11	0.25	.....
MnO	0.04	0.04	0.06	0.21	0.03	0.07	0.03	.....	.....
BaO	0.09	0.13	0.07	0.09	0.05	Tr.	0.01	.....	.....
SO_3	n.d.	Tr.	0.00	0.03	0.00	Tr.	Tr.	.....	.....
S	0.02	0.01	0.01	0.05	0.01	0.01	0.04	.....	.....
Total	100.07	100.15	99.91	99.83	99.70	99.93	100.11	99.48	100.29
Sp. gr. $f°/4°$	2.670	2.616	2.642	2.682	2.632	2.671	2.658	.....	.....

* Loss on ignition.

1. Granite gneiss, Cold Spring Granite Company quarry, Morton, Minnesota. S. S. Goldich, analyst.

2–7. Residual clays from Morton–Redwood Falls area; locations given in Fig. 1. See also the text. S. S. Goldich, analyst.

8. Hornblende-biotite-gneiss of Morton. *U.S. Geol. Surv. Bull. 157* (1899), p. 75. A. D. Meeds, analyst.

9. Decomposed gneiss from locality south of Morton. *U.S. Geol. Surv. Bull. 678* (1919), p. 217. F. F. Grout, analyst.

Petrography.—The essential minerals of the gneiss are oligoclase, orthoclase, microcline, quartz, biotite, and hornblende. Accessory minerals are magnetite, apatite, titanite, allanite, and zircon. Secondary products—sericite, chlorite, epidote, calcite, leucoxene, and

pyrite—are not abundant. The plagioclase is zoned irregularly, and the more calcic portions are sericitized. Irregularities in extinction suggest a lack of homogeneity which may have resulted from magmatic evolution and probably also from contamination.

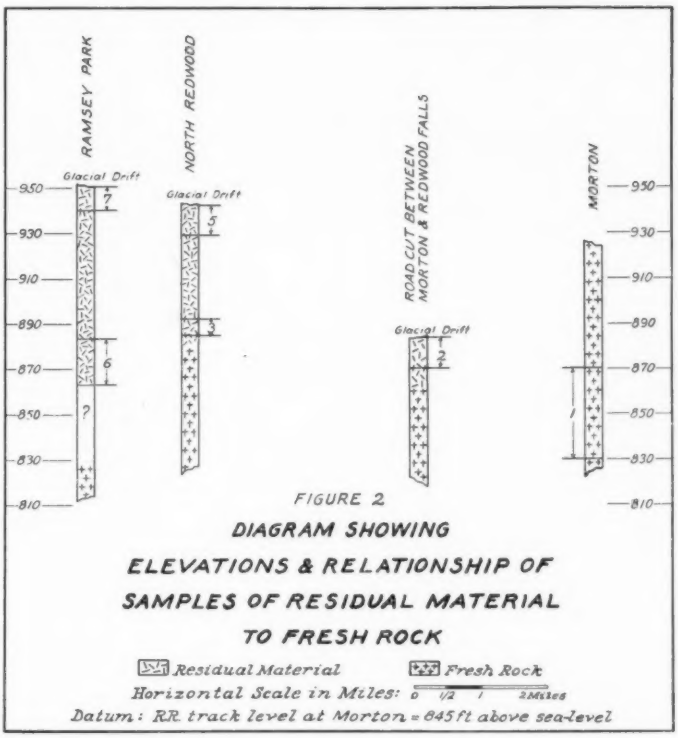

FIG. 2.—Diagram showing elevations and relationship of samples of residual material to fresh rock.

Thiel and Dutton[11] have classed the Morton rock as granite gneiss, but the calcic character of the gneiss should be noted. This is a result of reaction between the granitic magma and the minerals of the schist. The conversion of hornblende to biotite with the con-

[11] *Op. cit.*, p. 88.

sequent freeing of lime no doubt made the feldspar more calcic. The estimated mineralogical composition (Table 6) suggests a quartz monzonite. In Johannsen's classification, *adamellite* would be appropriate. In the C.I.P.W. system, the rock is *toscanose*.

RESIDUAL CLAYS

Field relationships.—Weathering in the Morton–Redwood Falls area has not proceeded uniformly. In the uppermost part of the

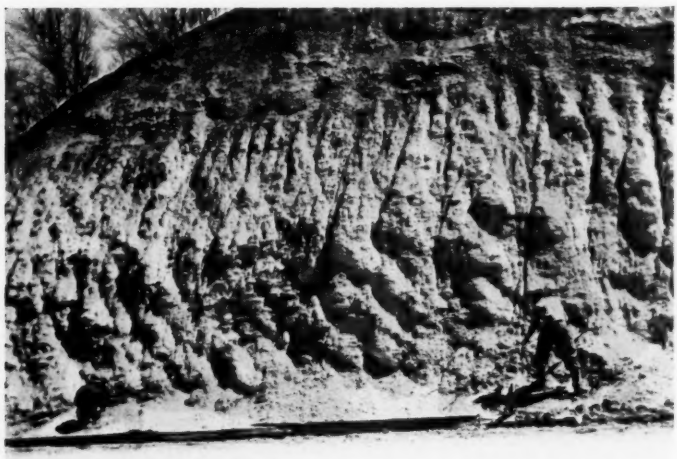

FIG. 3.—Residual clay along the highway between Redwood Falls and North Redwood.

quarry at Morton, weathering has a definite relationship to the jointing. Spheroidal weathering was probably initiated at considerable depth below the surface. In the exposures of residual clay less altered portions, frequently pegmatitic zones, protrude as ridges (Figs. 3 and 4).

In all the outcrops sampled some evidence of the nature of the original rock can be found in texture and in structures which have been retained. The less altered portions were of aid in determining the parent-gneiss. The abundance of quartz grains in the residual clay indicates a granite gneiss not very different from that exposed at Morton. Quartz veins, though somewhat shattered and broken,

can be traced vertically from the base to the top of certain outcrops, and these together with color banding inherited from the gneiss show the residual character of the clay.

FIG. 4.—Bluff of irregularly weathered gneiss in Ramsey State Park, Redwood Falls.

Location and description of samples.—The residual clays were sampled by channeling the exposures. The elevations given for the samples were taken with a Paulin surveying aneroid. The relationships are shown in Figure 2.

Sample No. 2: Elevation of base, 871 feet. Tan-colored residual clay in a road-cut on the south side of U.S. Highway No. 71 between

Morton and Redwood Falls was sampled in a 12-foot vertical section. Quartz grains and partially kaolinized feldspar are visible. The presence of knobs of fresh gneiss near by at an elevation of 850 feet and the lateral gradation of the residual clay in the road-cut into comparatively fresh gneiss suggest that the material represented in the sample is not far removed from bedrock.

Sample No. 3: Elevation of base, 885 feet. The sample was taken from an exposure along the highway between Redwood Falls and North Redwood about $\frac{1}{10}$ mile south of the intersection of the highway and the railroad tracks in North Redwood. The dry material is brown. Quartz and altered feldspar are conspicuous.

Sample No. 4: Dark-gray clay collected and described by Grout: "Between Redwood Falls and Morton, on the south bank of Minnesota River, a test pit has been dug by the Morton Brick and Tile Co. into a bank of residual decomposed gneiss, which has retained its granitoid texture."[12]

Sample No. 5: Elevation of base, 929 feet. The sample represents the upper 13 feet of about 25 feet of clay on the east side of the highway from Redwood Falls to North Redwood uphill from sample No. 3. The dry color is light gray. The base of the sample is about 38 feet above the top of sample No. 3.

Sample No. 6: Elevation of base, 863 feet. Light-green to gray clays extend to a depth exceeding 90 feet. The sample is from a 20-foot vertical channel, the base being a few feet above river level.

Sample No. 7: Elevation of base, 940 feet. Sample of 10 feet of yellow-brown clay from the top of bluff above sample No. 6. No definite structure is visible, and the clay appears uniform and compact.

Mineralogy of the residual clays.—The residual clays are composed chiefly of kaolinite and quartz, the latter derived from the original mineral of the gneiss. Other original minerals are not common. Besides the kaolinite and other clay minerals not positively identified, iron oxides, siderite, leucoxene, and a chlorite-like mineral are weathering products.

Quartz: The analyses in Table 1 are arranged in order of increasing water content, which might be taken as an indication of the ex-

[12] *Op. cit.*, p. 217.

tent of decomposition. It will be noted that an increase in water over that contained in the fresh gneiss is accompanied by an increase in alumina and, in general, by a decrease in silica. The series for silica, however, is not a regular one, and in order to determine to what extent the silica content is influenced by the amount of quartz present, quantitative study of the quartz was attempted by the method suggested by Knopf.[13] This method is based on the solubility of silicate minerals in hydrofluosilicic acid, which does not attack quartz appreciably. A definite correlation is found between the quartz and the silica contents of the samples; high silica in a clay sample reflects a high quartz content.

TABLE 2

QUARTZ CONTENT OF RESIDUAL CLAY SAMPLES*

	1	2	3	4	5	6	7
SiO_2	71.54	69.89	68.09	61.75	70.30	57.53	55.07
Quartz	28.4	35.0	38.7	31.7	41.9	26.3	24.9

* Estimated by the hydrofluosilicic acid method. The quartz of sample No. 1 is normative quartz. The silica percentages are from Table 1.

Clay mineral: The chief clay mineral formed by the decomposition of the silicates of the gneiss is kaolinite ($H_4Al_2Si_2O_9$). The composition of the clay mineral was first demonstrated by partial chemical analyses quoted in Table 3. More recent analyses of the clays in the lower part of the Cretaceous formation are included in the table. These clays were derived by the re-working of the weathering products of the gneiss in this region. A sample of the white clay from the Cretaceous collected by Gilman Berg was x-rayed by Professor J. W. Gruner, of the University of Minnesota, who reported the pattern as that of kaolinite.[14]

In the residual clays some nontronite, or iron-bearing clay mineral, may be present. A green micaceous mineral apparently derived from the biotite of the gneiss approaches kaolinite in composition.

Heavy minerals: Bromoform separations were made on the

[13] Adolph Knopf, "The Quantitative Determination of Quartz in Dusts," *U.S. Public Health Service Reports*, Vol. XLVIII, No. 8 (1933), pp. 183–90.

[14] Verbal communication from Mr. Berg, 1935.

washed samples of the fresh gneiss and the residual clays according to the procedure described. The heavy mineral analyses are given in Table 4. Much of the ilmenite in the clay samples is fresh, but grains in various stages of alteration are most abundant. Limonite,

TABLE 3

ANALYSES OF CRETACEOUS CLAYS AND CLAY WASHED FROM WEATHERED GNEISS

	1	2	3	4
SiO_2	43.86	45.92	45.14	44.12
Al_2O_3	41.82†	39.84†	37.94	38.39
Fe_2O_3	41.82†	39.84†	1.01	1.06
FeO	41.82†	39.84†	1.01	1.06
MgO	Very small	Very small	0.08	0.11
CaO			0.46	0.28
Na_2O	n.d.	n.d.	0.36	0.30
K_2O	n.d.	n.d.	0.09	0.17
H_2O*	14.65	14.12	14.10	14.70
TiO_2	.....	.....	1.09	1.17
Total	100.33	99.88	100.27	100.30

* H_2O by ignition.

† With the alumina of No. 1 is associated a trace of iron and with that of No. 2 a little iron and about 0.3 per cent of TiO_2.

1. Clay from Morton. *U.S. Geol. Surv. Water-Supply Paper 256* (1911), p. 310. F. F. Grout, analyst.
2. Clay washed from a decomposed granite, Redwood Falls. *Ibid.* F. F. Grout, analyst.
3. Smooth white clay from Cretaceous deposits near Morton. *U.S. Geol. Surv. Bull. 678* (1919), p. 219. Analysis by Bureau of Standards.
4. Concretionary clay from Cretaceous deposits near Morton. *Ibid.* Analysis by Bureau of Standards.

hematite, and leucoxene stained by secondary iron oxides could not be differentiated readily and were counted together. Similarly, biotite grades into a green mineral counted as chlorite.

Biotite in the gneiss as well as some of the unaltered mineral in certain clay samples gave beta = 1.640. The green micaceous mineral in the heavy mineral fractions has a lower refractive index. Further weathering of this mineral seems to leach it to such an extent

that it floats on bromoform. Flakes of a faint greenish color in the light mineral fractions gave beta = 1.570, and flakes of essentially

TABLE 4

HEAVY-MINERAL ANALYSES OF FRESH GNEISS AND RESIDUAL CLAYS IN THE MORTON–REDWOOD FALLS AREA

	1	2	3	4	5	6	7
Washed sample (percentage weight of original)	91.0	63.4	60.8	60.0	58.0	62.8	56.0
Heavy fraction (percentage weight of washed sample)	7.36	2.62	4.05	9.54	1.37	3.45	3.10
	Percentage approximate weight of total heavy minerals						
Minerals:							
Allanite	0.06	*	*	*	*	*	*
Apatite	1.91	.04	Tr.	0.01	Tr.	0.03	0.03
Biotite	76.89	18.55	0.70	29.20	0.68	Tr.	0.00
Chlorite	Present	Present	.....	Present	.....	11.10	4.89
Collophane(?)	None	0.33	0.00	0.01	0.01	?	?
Epidote	2.24	0.09	0.14	1.78	0.36	0.19	0.24
Hornblende	9.84	0.13	0.00	0.03	0.00	2.50	0.33
Ilmenite	2.27						
Leucoxene	0.11						
Limonite and hematite	0.01	79.73	98.34	19.55	97.05	70.14	86.00
Magnetite	5.65						
Pyrite	0.05	0.07	Tr.	0.02	0.01	0.03	0.29
Siderite	None	0.28	0.05	48.53	0.18	11.62	5.13
Titanite	0.49	0.00	0.00	0.01	0.01	0.00	0.00
Zircon	0.43	0.74	0.76	0.81	1.68	4.36	3.09
Unknown	0.05	0.04	0.01	0.05	0.02	0.03	0.00
Total	100.00	100.00	100.00	100.00	100.00	100.00	100.00

* Included with epidote.

the same appearance but colorless have an index of refraction of about 1.565 in basal section. This mineral is probably kaolinite. A

similar alteration of biotite to kaolinite has been described by Tullis and Laney,[15] who found that a green micaceous mineral represents an intermediate stage in the alteration. These men question the identity of this product with chlorite.

Light Minerals: Of particular interest is the feldspar content of the light fractions obtained from bromoform separations. Frequency percentages of the quartz and feldspar in these fractions are given in Table 5. Besides the quartz and feldspar of sample No. 1, a small amount of biotite and chlorite is listed as "other minerals." The feldspar of the residual clay samples is almost wholly composed of the potash feldspars, orthoclase and microcline. Kaolinite and the greenish, chlorite-like mineral constitute the remaining minerals.

TABLE 5

MINERALOGICAL ANALYSES OF LIGHT FRACTIONS OF WASHED SAMPLES

	1	2	3	4	5	6	7
Quartz	48	41	54	41	60	n.d.	n.d.
Feldspar	48	38	23	21	21	None	None
Others	4	21	23	38	19	n.d.	n.d.

ESTIMATED MINERALOGICAL COMPOSITION OF FRESH AND WEATHERED SAMPLES

The mineralogical compositions of the clay samples and of the fresh Morton gneiss were estimated from the results of the mineralogical study in conjunction with the chemical data. It is difficult to state the accuracy of the estimates in Table 6, but the different sources of information used in the compilation afforded some check on the reasonableness of the results. Thus a rough estimate of the potash feldspar can be made from the light-mineral analyses and can be checked by the potash available in the chemical analyses.

After allotting various oxides to the minerals known to be present in accordance with calculations similar to those employed in the C.I.P.W. system, the residual silica was allotted to quartz. In every case, the calculated quartz approached within 10 per cent of the values obtained in the quartz determinations listed in Table 2.

[15] E. L. Tullis and F. B. Laney, "The Composition and Origin of Certain Commercial Clays of Northern Idaho," *Econ. Geol.*, Vol. XXVIII (1933), p. 487.

CHEMICAL AND MINERALOGICAL CHANGES

The chemical and mineralogical changes resulting from the weathering of the gneiss in the Morton–Redwood Falls area are shown in a series of diagrams (Figs. 5, 6, 7, and 8). These will be considered at this time.

TABLE 6

MINERALOGICAL COMPOSITION OF THE MORTON GNEISS AND RESIDUAL CLAYS ESTIMATED FROM MINERAL AND CHEMICAL ANALYSES

	1	2	3	4	5	6	7
Quartz	30.0	35.0	40.0	31.0	43.0	28.0	25.0
Potash feldspar	19.0	31.0	18.0	19.0	13.0	2.0 (?)	1.0 (?)
Plagioclase	40.0	4.0	1.0	1.0	1.0	?	?
Biotite (Chlorite)	7.0	0.8	Tr.	4.0	Tr.	0.5	0.2
Hornblende	1.0	Tr.	None	Tr.	None	0.1	Tr.
Magnetite	1.0						
Ilmenite	0.5	3.0	5.0	4.0	2.0	6.0	6.0
Leucoxene and secondary iron oxides	Tr.						
Calcite	0.3	.....	.....	.....	.....	.....	.....
Siderite	.....	0.6	0.1	5.0	0.5	2.0	0.9
Titanite	0.04	None	None	Tr.	Tr.	None	None
Epidote	0.2	Tr.	Tr.	0.2	Tr.	Tr.	Tr.
Apatite	0.2	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.
Zircon	0.02	0.02	0.03	0.08	0.02	0.15	0.10
Allanite	Tr.	.....	.....	.....	.....	.....	.....
Kaolin	.....	26.0	36.0	35.0	40.0	59.0	66.0

"Straight-line" diagram.—Figure 5 is a "straight-line" diagram in which the residual clay samples have been compared with the fresh gneiss in the Morton quarry. If alumina is considered to have remained constant during the weathering process, the constituents of any one sample, which fall to the right of the point indicating alumina for that sample, have been lost; those to the left, gained. The pattern developed in the diagram indicates some degree of uniformity in the behavior of most of the constituents.

The progressive alterations shown by K_2O and BaO in Figure 5 and their close relationship are noteworthy as is the more erratic behavior of FeO and CO_2. Residual clays containing appreciable amounts of ferrous iron also have considerable carbon dioxide, and in these samples siderite is found. Siderite composes about 5 per

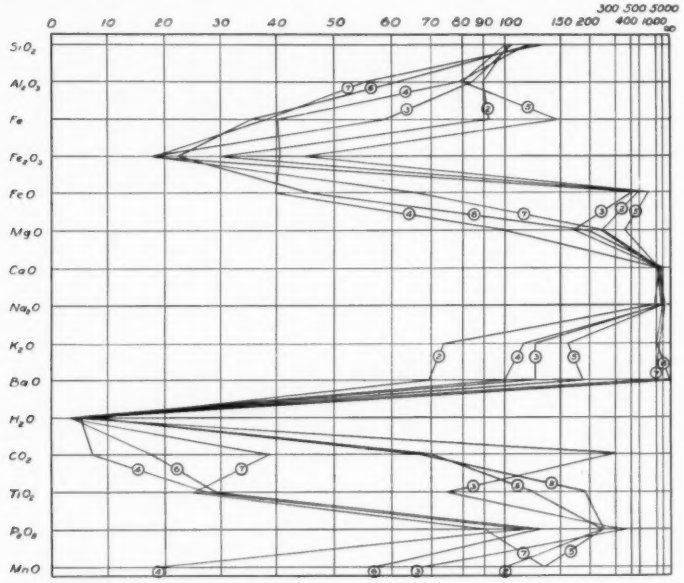

FIG. 5.—Representing the losses and gains of constituents by weathering of gneiss in the Morton–Redwood Falls area.

cent of sample No. 4. Under the usual conditions of weathering prevailing at the surface, ferrous iron is oxidized. The more or less hydrated ferric oxides are relatively stable, but, in the presence of reducing agents, iron may be taken into solution. Solution and transportation of iron as the bicarbonate apparently did take place in this region. Downward migration with subsequent precipitation as the carbonate may be in part the cause for the differences in iron content of the samples. There may have been differences in the original material also, and the behavior of TiO_2 and MnO is suggestive of such differences.

Variation diagrams.—Variation diagrams are useful in showing relationships in a rock series. In Figure 6 the variation curve for lime and soda each follows that of plagioclase. The oligoclase of the

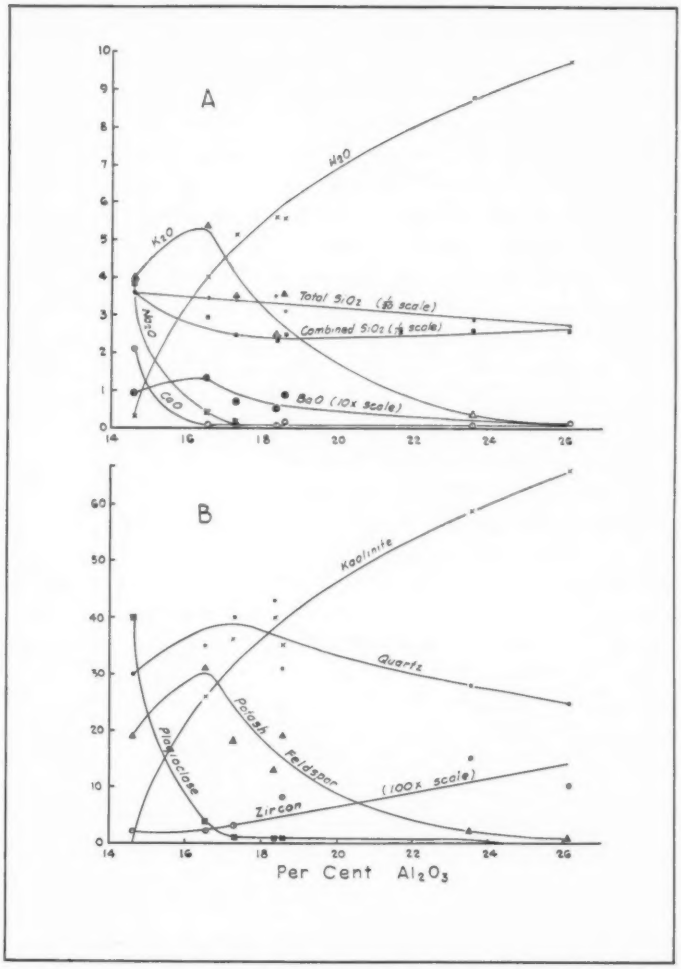

FIG. 6.—Variation diagrams of the series of fresh and decomposed granite gneiss samples from the Morton–Redwood Falls area.

gneiss is readily attacked, and the lime, soda, and some of the silica are removed. In contrast with these curves, potash and potash feldspar curves show an upward trend in the early stages of weathering. The feldspar of the light-mineral fractions is largely potash feldspar, and it is believed that the early destruction of the plagioclase results in a mechanical concentration of the orthoclase and microcline. Thus in sample No. 2, potash feldspar is estimated to compose about 35 per cent of the clay.

The behavior of baria is of interest in connection with the alteration of the feldspar. Baria is usually considered to be closely related to lime, and rules for calculations involving chemical analyses commonly call for the addition of baria to the lime. Inspection of the chemical analyses, however, reveals no correlation between the amounts of lime and baria, but rather one between potash and baria. This is clearly shown in Figure 5. The presence of baria in some potash feldspars is well known. The barium molecule, $BaAl_2Si_2O_8$, celsian, enters into solid solution with the orthoclase molecule, $KAlSi_3O_8$. In the series of samples from the Morton–Redwood Falls area some of the potash must be assigned to biotite. Biotite, however, is not abundant, and in the residual clays the potash and baria can be correlated with the potash feldspar content of the clays. Only when the orthoclase and microcline have been thoroughly decomposed are the potash and baria largely removed. Such a condition of profound leaching has been attained in samples No. 6 and No. 7, and the chemical analyses of these samples reveal only small amounts of potash and traces of baria.

It has been suggested that the decomposition of the plagioclase in the early stages of weathering results in a mechanical concentration of the more resistant orthoclase and microcline. Similarly, the decomposition of the feldspar causes an apparent increase in the quartz content of the clay samples, but the high point in the quartz curve beyond which it declines comes at a later stage in the weathering process. The total silica curve shows a decline along an essentially straight line. Actually, this curve is a composite of the "combined" silica curve (Fig. 6*A*) and of a quartz curve such as the one shown in Figure 6*B*. The decomposition of the feldspar with the loss of silica is reflected in the downward slope of the combined silica

curve, but at the same time the quartz curve is ascending, compensating for the loss of silica from the destruction of silicates. Finally, after more or less complete alteration of the original aluminum silicates to the stable product, kaolinite, quartz is slowly dissolved, and both the curve for quartz and that for total silica have a downward slope.

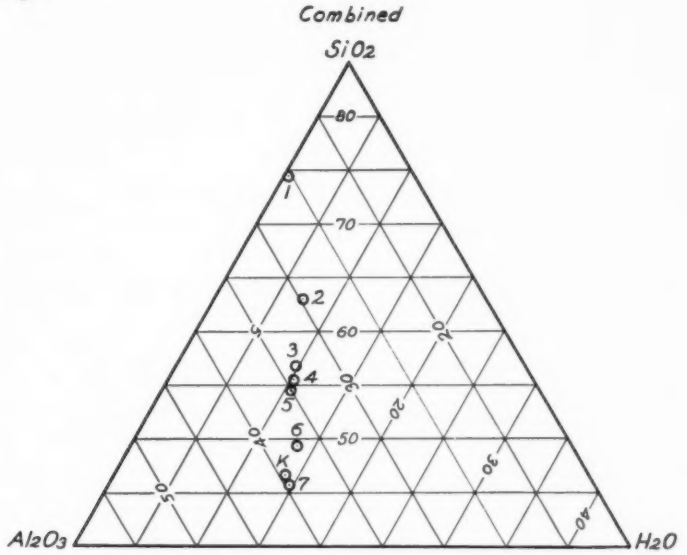

FIG. 7.—Diagram showing gradation from unweathered aluminum silicates in gneiss (1) to kaolinite (*K*) in residual clay samples (2–7).

Kaolinization diagram.—The progress of the decomposition of the silicates to kaolinite is shown in Figure 7. This diagram is similar to one used by Mead[16] to represent the gradation from syenite to bauxite in the Arkansas deposits. In the present study it was found that plotting the percentages of silica, alumina, and combined water was ineffective in showing the extent of alteration because the quartz contributes heavily to the total silica of the clays, masking the constituents of interest which are combined silica, alumina, and com-

[16] W. J. Mead, "Occurrence and Origin of the Bauxite Deposits of Arkansas," *Econ. Geol.*, Vol. X (1915), p. 49.

bined water. The quartz percentages (Table 2), therefore, were subtracted from the figures for silica to obtain the percentages of combined silica. The combined silica, the alumina, and the combined water were recalculated to 100 and plotted with the result that the series of samples shows a progression toward the theoretical composition of kaolinite.

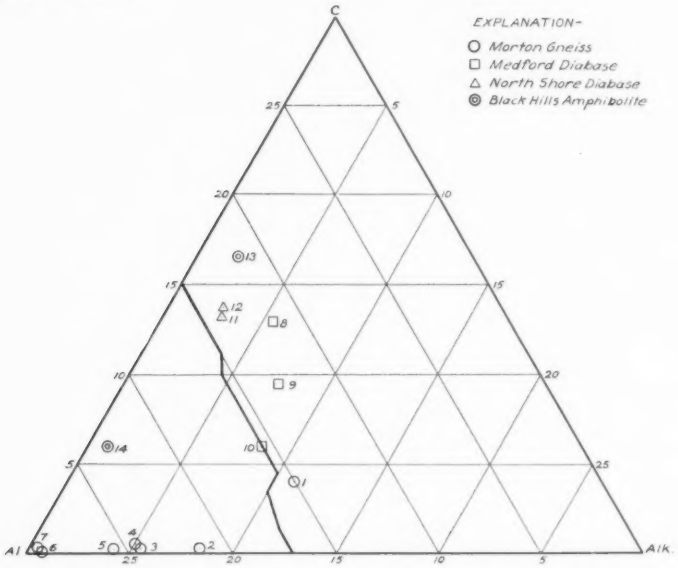

FIG. 8.—Fresh and weathered rocks plotted on the *Al-C-Alk* diagram (after Osann). Rocks with sedimentary characteristics fall to the left of the heavy line; igneous rocks, to the right.

Sample No. 7 falls beyond the point representing kaolinite, indicating an excess of alumina and water. This may be caused by errors in the analytical determinations or in the estimation of quartz, or it may indicate the presence of an aluminous hydrate. Grout[17] records a test by Bowles indicating 3–4 per cent of bauxite in some of the clays of this area. The clays tested by Bowles, however, were taken from the overlying Cretaceous deposits and show a concretionary texture. None of the residual clays sampled exhibited this texture.

[17] *Op. cit.*, p. 219.

An excess of water may be explained by the presence of hydrated oxides of iron.

"Al-C-Alk" diagram.—The rapid impoverishment in lime and in soda and the more gradual loss of potash are the most striking features of the decomposition of the gneiss. The progress of the alteration is well displayed in the *Al-C-Alk* diagram of Figure 8. In this figure the co-ordinates are Al_2O_3, $CaO + BaO$, and $Na_2O + K_2O$. The *Al* corner of the triangle, therefore, would correspond to kaolinite.

DISCUSSION AND SUMMARY

Certain discrepancies or irregularities, both in chemical and in mineralogical data obtained in the study of the fresh granite gneiss and of its residual weathering products, are undoubtedly the result of differences in parent-material. These differences have been partly smoothed out in the curves of Figure 6 and in the diagram of Figure 7. If the data are generalized in this manner, the picture of the weathering processes in the Morton–Redwood Falls area is believed to be fairly accurate.

Besides differences in parent-rock which might be reflected in the composition of the residual clays, other factors tend to complicate the data. The formation or the introduction of secondary heavy minerals, such as siderite, affects the percentages of the original minerals in the heavy concentrates. However, it will be noted (Table 4) that zircon constitutes a larger percentage of the total heavy minerals in the clay samples than in the fresh rock. The obvious conclusion is that the heavy minerals of the gneiss have been largely destroyed, and those minerals remaining represent in some degree the relative resistance or stability of the minerals. Of the heavy minerals, hornblende, epidote, titanite, and apatite appear to be least stable. Ilmenite-magnetite seems to be more resistant, and zircon, most stable. Even the zircon, however, shows signs of attack in its dusty, granular appearance. The altered mineral is brown, has a lower birefringence than fresh zircon, and tends to become opaque.

If alumina is considered constant, the constituents lost, in order of rate of loss, are Na_2O, CaO, MgO, K_2O, BaO, P_2O_5, SiO_2, S, and FeO. Those gained are H_2O, CO_2, Fe_2O_3, TiO_2, Fe, and MnO.

The following order of stability is suggested for the original minerals of the gneiss:

Plagioclase, epidote, hornblende, titanite, apatite	Least stable
Ilmenite-magnetite, Biotite, Microcline, orthoclase	Moderately stable
Zircon, quartz	Most stable

The abundant minerals in the residual clays are kaolinite and quartz. The absolute behavior of the quartz during weathering is not certain, but, after liberal allowances have been made for possible differences in parent-rock and for errors inherent in the technique, it appears that some of the quartz has been removed by solution. Ross and Kerr[18] cite examples of the replacement of quartz by kaolinite.

THE MEDFORD DIABASE

GENERAL STATEMENT

The Medford diabase in eastern Massachusetts has afforded material for a number of investigations primarily concerned with the weathering of this rock. Probably the best known is the classical study by Merrill.[19]

More recent papers are concerned with the time of weathering. Wolf[20] and Lane [21] hold that the weathering is mainly older than the Wisconsin glaciation as has been suggested by Wilson.[22] Billings and Roy[23] believe that the weathering is younger than the last ice inva-

[18] C. S. Ross and P. F. Kerr, "The Kaolin Minerals," *U.S. Geol. Surv. Prof. Paper 165-E* (1931), p. 174.

[19] G. P. Merrill, "Disintegration and Decomposition of Diabase at Medford, Massachusetts," *Geol. Soc. Amer. Bull. 7* (1896), pp. 349–62; *Rocks, Rock-weathering, and Soils* (New York: Macmillan Co., 1913), pp. 198–203.

[20] Arthur Wolf, "Weathering of the Medford Diabase—Pre- or Postglacial?" *Jour. Geol.*, Vol. XL (1932), pp. 459–66.

[21] A. C. Lane and Arthur Wolf, "Reply to Discussion of 'Weathering of the Medford Diabase—Pre- or Postglacial?' " *Jour. Geol.*, Vol. XLI (1933), pp. 661–66.

[22] A. W. G. Wilson, "The Medford Dike Area," *Boston Soc. Nat. Hist.*, Vol. XXX (1901), pp. 353–74.

[23] M. P. Billings and C. J. Roy, "Weathering of the Medford Diabase—Pre- or Postglacial?" *Jour. Geol.*, Vol. XLI (1933), pp. 654–61.

sion. They are supported by the earlier work of Crosby,[24] of Barton,[25] and of Merrill.[26] In this paper new chemical data obtained on a series of samples generously supplied by Dr. A. C. Lane are presented. Descriptions of sample locations are given in Table 7.

PETROGRAPHY

Petrographic descriptions of the Medford diabase have been given by Hobbs,[27] by Jaggar,[28] and by Lane.[29] The notes which follow were made from study of thin sections of fresh rock No. 9. Plagioclase, hornblende, augite, magnetite, biotite, and apatite are the

TABLE 7

LOCATIONS OF SAMPLES AS RECEIVED FROM A. C. LANE

Sample No.	Lane's No.	Description
8.....	.....	Fresh Medford diabase from an abandoned quarry
9.....	.7562	Relatively fresh diabase. Opposite 93 Governor's Ave., Medford, Mass. Lat. 42°25′; Long. 71°6′54″
10.....	.7561	Weathered Medford diabase. Four feet from surface in disintegrating crack. Location as of sample No. 9
.7558.....	.7558	Decayed diabase, Cousens Gymnasium, Tufts College
.7559.....	.7559	Glacial till overlying .7558. Cousens Gymnasium, Tufts College
.7560.....	.7560	Residual boulder of diabase. Cousens Gymnasium, Tufts College

essential minerals. Quartz, orthoclase, epidote, chlorite, calcite, and pyrite make up the rest of the rock. The plagioclase is zoned labradorite partly replaced by orthoclase and by intergrowths of orthoclase and quartz. The plagioclase has been much altered, and individual crystals have been completely changed to a felt of sericite

[24] W. O. Crosby, "Contributions to the Geology of Eastern Massachusetts," *Boston Soc. Nat. Hist.*, Vol. III (1880).

[25] G. H. Barton, "Bowlders Formed *in Situ*," *Tech. Quarter.*, Vol. V (1892), pp. 401–5.

[26] "Disintegration and Decomposition of Diabase at Medford, Massachusetts," *loc. cit.*, p. 358.

[27] W. H. Hobbs, "On the Petrographical Characters of a Dike of Diabase in the Boston Basin," *Mus. Comp. Zoöl. Bull. 16* (1888), pp. 1–13.

[28] T. A. Jaggar, "An Occurrence of Acid Pegmatyte in Diabase," *Amer. Geol.*, Vol. XXI (1898), pp. 203–13.

[29] A. C. Lane, "Die Korngrösse der Auvergnosen," *Rosenbusch-Festschrift* (1906).

fibers. Pyroxene is altered to hornblende. Reddish-brown biotite forms rims around grains of magnetite. Apatite, which is rather

TABLE 8

CHEMICAL ANALYSES OF FRESH AND WEATHERED DIABASE FROM MEDFORD, MASSACHUSETTS

	Analyst								
	S. S. Goldich (1935)			C. N. Whitney (1911)*			G. P. Merrill (1896)†		R. C. Sweetser (1887)‡
	Fresh		Weathered	Fresh	Weathered		Fresh	Weathered	Fresh
	8	9	10	6	5	2	1	3	
SiO_2	47.59	51.44	51.38	50.30	49.02	50.58	47.28	44.44	48.75
Al_2O_3	14.78	15.67	15.92	17.32	19.11	17.93	20.22	23.19	17.97
Fe_2O_3	4.39	2.25	10.92	4.07	8.74	8.87	3.66	12.70	0.41
FeO	9.11	8.37	2.60	10.65	6.30	5.46	8.89	.....	13.62
MgO	4.28	2.09	1.26	3.94	3.05	2.56	3.17	2.82	3.39
CaO	9.06	6.31	3.52	5.25	4.66	3.27	7.09	6.03	8.82
Na_2O	3.36	4.42	4.16	3.96	4.07	4.55	3.94	3.93	1.63
K_2O	1.18	1.91	1.57	1.54	1.02	1.21	2.16	1.75	2.40
H_2O+	1.35	1.67	2.94	1.69	3.26	2.95	2.73§	3.73§	0.60
H_2O-	.24	0.41	1.70	.....	.....	.....	.....	.....	.....
TiO_2	3.55	1.95	2.52	n.d.	n.d.	n.d.	n.d.	n.d.	0.99
P_2O_5	0.42	0.97	1.03	0.20	0.54	0.70	0.68	0.70	0.68
MnO	0.17	0.18	0.22	0.61	0.45	0.37	0.77	0.52	0.91
CO_2	0.65	2.14	0.07	0.37	None	None	.....	.....	Tr.
S	0.13	0.12	0.01	Tr.	None	None	.....	.....	Tr.‖
Cl	.....	.....	.....	Tr.	Tr.	Tr.	.....	.....	.....
	100.26	99.90	99.82	99.90	101.22¶	98.45	100.59	99.81	100.17
Less O≡S	0.05	0.05	.....	.....	.....	.....	.....	.....	.....
Total	100.21	99.85	99.82	99.90	100.22	98.45	100.59	99.81	100.17
Sp. gr.	2.945	2.80	2.749	2.87	2.74	.....	.....	.....	2.985

* Published by Arthur Wolf, *Jour. Geol.*, Vol. XL (1932), p. 460.

† *Geol. Soc. Amer. Bull.* 7 (1896), p. 353.

‡ Published by W. H. Hobbs, *Mus. Comp. Zoöl. Bull. 16* (1888), p. 9.

§ Loss on ignition.

‖ FeS_2.

¶ Whitney's error in summation.

abundant, appears to be corroded. The changes noted are probably in part deuteric, but the sericite, chlorite, calcite, and pyrite are hydrothermal.

CHEMICAL ANALYSES

Chemical analyses of three of the samples collected by Lane are given in Table 8. The published analyses of fresh and weathered diabase are also included. Of the new analyses, samples No. 9 and No. 10 are from the same locality as Whitney's[30] material (Nos. 6, 5, 2). Sample No. 8 represents a more basic facies of the diabase and seems to be much like the material studied by Merrill. Samples No. 8 and No. 9 reveal differences in the Medford diabase, a fact known and referred to by previous writers. When classified in the C.I.P.W. system, these samples fall in different classes (Table 9).

TABLE 9

	Sample No. 8	Sample No. 9
Class:	(II) III, salfemic	II (III), dosalic
Subclass:	1	1
Order:	5, perfelic	5, perfelic
Rang:	3, alkalicalcic	2, domalkalic
Subrang:	4, dosodic	4, dosodic
Name:	CAMPTONOSE	AKEROSE

Assuming alumina to have remained constant, Merrill calculated the percentage of each constituent lost during weathering. Billings and Roy made similar calculations using Whitney's analyses, and the writer has done the same for analyses No. 9 and No. 10. Inasmuch as the quality of the published analyses[31] has been mentioned by Billings, Roy, Lane, and Wolf, it is of interest to compare the calculations based on the new and old chemical analyses (Table 10).

The most obvious inconsistency in the data of Table 10 is in the figures for manganous oxide. A loss of 40 per cent of the MnO is indicated by the older analyses, but the new data show a gain for this constituent. Procedures formerly used in the determination of MnO commonly resulted in large positive errors. The modern colorimetric method using potassium periodate as the oxidizing agent is effective and leaves little to be desired for accuracy. Disregarding MnO, the calculated percentage losses are fairly consistent, and differences in samples of the weathered as well as of the fresh diabase may account for some of the discrepancies.

[30] C. N. Whitney, "The Medford Diabase Dike" (Tufts College thesis [1911]).

[31] The analyses by Sweetser and Merrill are rated as inferior by Washington, "Chemical Analyses of Igneous Rocks," *U.S. Geol. Surv. Prof. Paper 99* (1917), p. 1065.

The relative gains and losses based on analyses No. 9 and No. 10 are shown in the "straight-line" diagram, Figure 9. Samples 8, 9,

TABLE 10

PERCENTAGE OF EACH CONSTITUENT LOST IN WEATHERING OF MEDFORD DIABASE

(Alumina Considered Constant)

	A	B	C	D
SiO_2	18.03	11.74	2.88	1.68
Al_2O_3	0.00	0.00	0.00	0.00
Fe_2O_3	18.10	Gain	Gain	Gain
FeO		46.42	50.48	69.42
MgO	21.70	29.89	37.26	39.28
CaO	25.89	19.61	39.87	45.09
Na_2O	12.83	6.90	Gain	7.38
K_2O	29.15	40.01	24.10	19.09
H_2O+	Gain	Gain	Gain	Gain
H_2O-				
TiO_2	n.d.	n.d.	n.d.	Gain
P_2O_5	11.39	Gain	Gain	Gain
MnO	41.57	33.19	41.41	Gain
CO_2	n.d.	n.d.	n.d.	96.78
S	n.d.	n.d.	n.d.	92.80

A. Fresh and weathered by G. P. Merrill.
B. Fresh rock No. 6 and weathered rock No. 5, by C. N. Whitney.
C. Fresh rock No. 6 and weathered rock No. 2, by C. N. Whitney.
D. Fresh rock No. 9 and weathered rock No. 10, by S. S. Goldich.

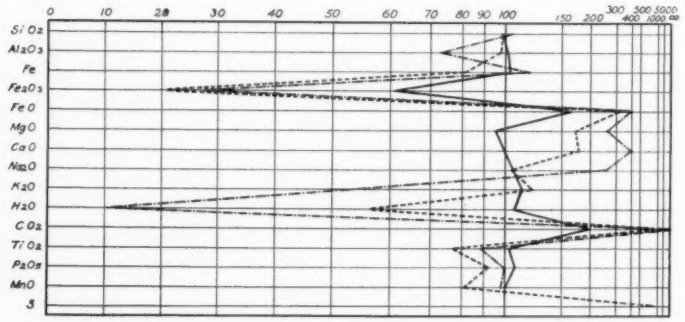

FIG. 9.—Representing the losses and gains of constituents by weathering of Medford diabase (dash line), North Shore diabase (solid line), and Black Hills amphibolite (dot-dash line).

and 10 have been plotted on the *Al-C-Alk* diagram (Fig. 6). Differences in the fresh rock can be seen in the positions of samples No. 8 and No. 9. The changes effected by weathering are shown by the differences in the positions of weathered sample No. 10 and the fresh rock No. 9. The weathered diabase is on the boundary line as drawn by Osann to differentiate sedimentary from igneous rocks.

HEAVY-MINERAL ANALYSES

The heavy minerals of the fresh and weathered diabase were separated, and the approximate percentages of the minerals in the heavy fractions of the washed samples are given in Table 11. An absolute loss in the heavy minerals as a result of weathering is indicated. The relative abundance of the heavy minerals in fresh and weathered diabase suggests also that the rate of decomposition is different for the various minerals. The order of stability, from least to most stable, appears to be chlorite, pyroxene, epidote, hornblende, biotite, apatite, magnetite-ilmenite.

DISCUSSION AND SUMMARY

It is not proposed to settle the question of the time of weathering of the Medford diabase in this paper. However, some of the data may be pertinent to the problem. In Table 12 ferrous and ferric iron determinations on the samples comprising the series received from A. C. Lane are given. As oxidation of the ferrous iron is characteristic of the weathering of the diabase, the percentages of the total iron in the bivalent (Fe^{++}) and in the trivalent (Fe^{+++}) states have been computed.

In the fresh diabase from 70 to 80 per cent of the total iron is bivalent. This range will include earlier analyses not shown in Table 12. In samples of weathered diabase, depending on the degree of alteration, a smaller fraction of the total iron is ferrous—from 20 to 68 per cent. In the sample of glacial till 59.4 per cent of the total iron is bivalent, but in weathered diabase immediately below this till only 37.5 per cent of the total iron is bivalent. It is not known what percentage of the iron was bivalent in the original fresh till, but it is most unlikely that it was as high as that of the diabase.

If it is assumed that the rate of oxidation of ferrous iron in the till and in the diabase is about the same, there has been less oxidation

(weathering?) of the till than of the diabase. This conclusion would be in accord with the hypothesis of preglacial weathering. The assumption of the rate of oxidation, however, may not be justified.

TABLE 11

HEAVY-MINERAL ANALYSES

	Sample No. 9 (Fresh)	Sample No. 10 (Weathered)
	Percentage weight	
Loss on washing	17.2	21.4
Heavy minerals in washed sample	22.5	15.9
	Percentage approximate weight of total heavy minerals	
Mineral:		
Hornblende	36.0	23.0
Chlorite	18.0	0.28
Biotite	14.0	9.4
Magnetite } Ilmenite }	12.0	24.0
Augite	10.0	3.1
Apatite	7.7	10.0
Epidote	1.3	0.51
Secondary iron oxides and leucoxene	0.90	29.0
Pyrite	0.22	0.08
Unknown	0.47	0.12

It is also true that some minerals containing ferrous iron are more stable than others, and possibly the minerals of the till are less susceptible to weathering than those of the diabase. On the other hand, permeability may be an important factor, favoring a higher rate of oxidation for the glacial till.

A single chemical criterion cannot always be applied safely, especially one requiring somewhat uncertain assumptions. The data, however, do indicate that considerable decomposition, involving oxidation and solution, has taken place locally, at least, in the weathering of the diabase. The extent of this decomposition will be referred to again in the following section dealing with the weathering of a similar rock.

TABLE 12

FERROUS AND FERRIC IRON IN SAMPLES OF MEDFORD DIABASE AND IN GLACIAL TILL

	8	9	10	.7558	.7559	.7560
Total iron (as Fe_2O_3)	14.51	11.55	13.81	12.04	5.61	12.05
Fe_2O_3	4.39	2.25	10.92	7.53	2.28	4.61
FeO	9.11	8.37	2.60	4.06	3.00	6.69
	Percentage of Fe^{++} and Fe^{+++} of total Fe					
Fe^{++}	69.8	80.5	20.9	37.5	59.4	61.7
Fe^{+++}	30.2	19.5	79.1	62.5	40.6	38.3

8. Fresh Medford diabase, from abandoned quarry.
9. Relatively fresh diabase, opposite 93 Governor's Ave., Medford, Mass.
10. Weathered diabase, opposite 93 Governor's Ave., Medford, Mass.
.7558. Weathered diabase, Cousens Gymnasium, Tufts College.
.7559. Glacial till overlying .7558.
.7560. Residual boulder of diabase, Cousens Gymnasium, Tufts College.

DIABASE ON THE NORTH SHORE OF LAKE SUPERIOR

GENERAL STATEMENT

Diabase Hill is situated about 7 miles northeast of Two Harbors opposite Encampment Island in Lake Superior. The diabase from which the hill is named is deeply weathered, and on the northern side considerable material has been taken out with a steam shovel for use as road metal. Brief mention of the weathering of this rock is made by Schwartz[32] and in an earlier work by Irving.[33]

[32] G. M. Schwartz, "A Guidebook to Minnesota Trunk Highway No. 1," *Minn. Geol. Surv. Bull. 20* (1925), p. 71.

[33] R. D. Irving, "The Copper-bearing Rocks of Lake Superior," *U.S. Geol. Surv. Mon. 5* (1883), p. 286.

A portion of Diabase Hill was blasted out in the construction of the highway, exposing the unaltered rock which is fine to medium in grain. Green feldspar and magnetite are the only minerals readily visible in hand specimens. The diabase weathers to a nodular or pebbly coarse, yellowish to reddish-brown sand. The larger lumps generally have a core of comparatively fresh diabase. Sample No. 11 represents the fresh rock. Sample No. 12 was taken from the weathered material about 35 feet above the fresh sample.

TABLE 13

CHEMICAL ANALYSES OF FRESH AND WEATHERED DIABASE FROM NORTH SHORE, LAKE SUPERIOR

	11	12
SiO_2	46.88	46.91
Al_2O_3	20.98	20.27
Fe_2O_3	3.32	5.30
FeO	5.56	3.44
MgO	4.74	4.89
CaO	11.15	11.19
Na_2O	2.49	2.31
K_2O	0.29	0.25
H_2O+	1.45	1.35
H_2O-	1.59	2.52
CO_2	0.12	0.06
TiO_2	1.11	1.08
P_2O_5	0.15	0.14
MnO	0.12	0.12
S	0.02	0.02
Total	99.97	99.85
Sp. gr. $t°/4°$	2.791	2.756

11. Fresh diabase.

12. Weathered diabase. S. S. Goldich, analyst.

PETROGRAPHY

The fabric of the rock is diabasic and porphyritic. The primary essential minerals are plagioclase, augite, and magnetite-ilmenite. The plagioclase is zoned, and inclusions of magnetite and pyroxene are numerous in the outer margins. The central, more basic portions have a maximum symmetrical extinction angle of 38°, indicating $Ab_{32}An_{68}$. The presence of serpentine in the characteristic forms resulting from the alteration of olivine indicates that the latter was

an original mineral, though none was noted in thin sections. Other minerals, deuteric or hydrothermal rather than weathered, include biotite, uralite, chlorite, pyrite, and zeolite(?).

CHEMICAL ANALYSES

Chemical analyses of fresh and weathered rock (Table 13) are remarkably similar. The significant differences are in moisture content and in the proportions of ferrous and ferric oxides.

TABLE 14

HEAVY-MINERAL ANALYSES

	Sample No. 11 (Fresh)	Sample No. 12 (Weathered)
	Percentage weight	
Loss on washing	20.4	11.6
Heavy minerals in washed sample	24.5	23.7
	Percentage approximate weight of total heavy minerals	
Mineral:		
Pyroxene	78.0	74.0
Magnetite } Ilmenite }	19.0	24.0
Biotite } Chlorite }	2.5	0.90
Pyrite	0.15	0.15
Amphibole	0.10	0.04
Apatite	0.02	0.02

HEAVY-MINERAL ANALYSES

The fresh diabase and its residual weathering products were ground to pass a 60-mesh sieve, and the washed samples were subjected to the heavy-mineral technique. The mineralogical analyses of Table 14 suggest a slight concentration of magnetite and ilmenite

at the expense of pyroxene, but the differences are not great and approach the limits of accuracy of the method.

SUMMARY

The weathering of the diabase in northern Minnesota has not produced profound mineralogical changes. As in the case of the diabase from Medford, there was notable oxidation of the ferrous iron. The depth of weathering in both areas reaches to some 40 feet locally, but the significant feature is the fact that decomposition was important in the Medford occurrence and of slight importance in the weathering of the Minnesota rock.

The amount of decomposition suffered by the diabase in northern Minnesota is what the writer would expect as a postglacial effect in that region. It is believed that in the early stages of weathering, a small amount of chemical action is effective in bringing about a striking loss in coherence and a pronounced discoloration. The results of this stage of weathering are deceptive as has been demonstrated by Blackwelder.[34]

BLACK HILLS AMPHIBOLITE

GENERAL STATEMENT

A cut on the Northwestern Railroad track exposes amphibolite which occurs as a lens in a fine-grained sericite schist about $\frac{1}{4}$ mile northeast of Lead, South Dakota. The fresh rock grades upward through a weathered zone 15 feet thick, passing into soil. The outcrop is about 50 feet long and 20 feet wide. Samples No. 13, representing the fresh rock, and No. 14, the residual material, were collected by Dr. G. M. Schwartz, of the University of Minnesota.

The fresh phase is a massive, coarsely crystalline rock composed largely of dark-green hornblende. It weathers to a yellowish-brown silt containing coarse material up to about the size of gravel. The larger fragments on being broken display a fibrous structure and green color.

PETROGRAPHY

The amphibolite is composed chiefly of hornblende. Garnet, titanite, quartz, chlorite, calcite, pyrite, and a black opaque mineral,

[34] Eliot Blackwelder, "Exfoliation as a Phase of Rock Weathering," *Jour. Geol.*, Vol. XXXIII (1925), pp. 793–806.

possibly ilmenite, constitute about 25 per cent of the rock. The texture is apparently a recrystallized one, and the large metacrysts of hornblende include numerous small grains of other minerals. A linear arrangement of the small anhedral grains of quartz and ilmenite is particularly striking. The thin bands of the black opaque mineral are conspicuous in thin section and appear to be entirely independent of the minerals in which they are contained. The arrangement suggests a texture inherited from a pre-existing rock, possibly a schist (Fig. 10).

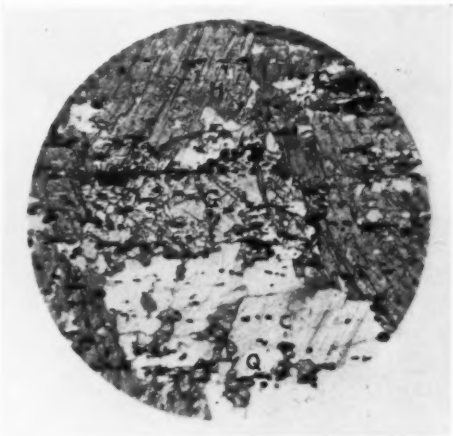

FIG. 10.—Parallel bands of ilmenite in Black Hills amphibolite suggesting a pre-existing structure independent of the present hornblende (*H*), garnet (*G*), and calcite (*C*). Some sections show a similar arrangement of small quartz (*Q*) grains. Ordinary light. ×30.

CHEMICAL ANALYSES

Chemical analyses of the fresh and weathered amphibolite are given in Table 15. The high FeO, MnO, TiO_2, and CO_2 content of the fresh rock are noteworthy. In the residual material, the abundance of water (13 per cent) is striking. Of the total water, more than half is lost on drying the sample below 110° C. This water is apparently adsorbed water. It is regained on allowing the dried sample to stand exposed to the air of the laboratory.

The relative gains and losses resulting from the weathering of this

rock are shown in Figure 9. The values plotted in this figure are based on analyses in Table 15. If the chemical data were recalculated on a moisture-free basis, 20 rather than 11 grams of the weathered rock would be required to furnish the water present in 100 grams of fresh amphibolite. The points on the diagram representing other constituents, excepting CO_2, would be shifted in the opposite direction.

TABLE 15

CHEMICAL ANALYSES OF FRESH AND WEATHERED AMPHIBOLITE FROM BLACK HILLS

	13	14
SiO_2	41.98	40.15
Al_2O_3	13.99	19.52
Fe_2O_3	2.22	11.97
FeO	15.64	3.95
MgO	5.45	2.09
CaO	10.91	2.73
Na_2O	1.23	0.45
K_2O	0.18	0.33
H_2O+	1.37	6.25
H_2O-	0.12	6.94
CO_2	1.43	0.00
TiO_2	3.84	4.34
P_2O_5	0.36	0.36
MnO	1.03	1.06
BaO	0.00	n.d.
S	0.07	n.d.
Total	99.82	100.14
Sp. gr. $f°/4°$	3.239	2.684

13. Fresh amphibolite.
14. Weathered amphibolite. S. S. Goldich, analyst.

If alumina is considered constant, the constituents lost, in order of decreasing percentage of loss, are CO_2, CaO, FeO, Na_2O, MgO, Fe, SiO_2, P_2O_5, MnO, and TiO_2. Those gained include H_2O, Fe_2O_3, and K_2O. Iron rather than alumina has been considered the invariable constituent in calculations of gains and losses of some occurrences.[35] On this basis, the constituents lost are CO_2, CaO, FeO, Na_2O, and MgO. Those gained are H_2O, Fe_2O_3, K_2O, Al_2O_3, TiO_2, MnO, P_2O_5, and SiO_2, water being greatest, and silica least.

[35] Merrill, *Rocks, Rock-weathering, and Soils*, p. 187.

The fresh amphibolite and its residual weathering product have been plotted on the *Al-C-Alk* diagram (Fig. 8). The pronounced chemical changes resulting from the decomposition of the rock can be seen in the positions of samples No. 14 and No. 13.

The position of the fresh sample in the "igneous rock field" might lend support to Paige's[36] contention that the amphibolites of this region were derived from igneous rocks. However, Runner,[37] although admitting the composition of the amphibolites approaches that of basic igneous rocks, holds, on the basis of field evidence, that they are of sedimentary derivation. He believes that calcareous, magnesian, and ferruginous shales by recrystallization under pressure yielded the amphibolites. The intrusive relationships of the dikelike masses in the schist country rock are said to be due to plastic flow under pressure without fusion of the rock minerals.

HEAVY-MINERAL ANALYSES

Samples of the fresh and weathered amphibolite were ground to pass a 60-mesh sieve. The washed samples were separated with bromoform, and the approximate weight percentages of the minerals in the heavy fractions are given in Table 16. The amount of heavy minerals obtained from the residual material is considerably less than that from the fresh amphibolite, and the relative abundance of the minerals is very different. This change is the result of the decomposition of the hornblende, which makes up 77 per cent of the heavy minerals of the fresh rock as opposed to about 7 per cent for the weathered material.

The light-mineral fraction from the fresh amphibolite was quartz and calcite. Quartz but no calcite was found in the light fraction of the weathered phase. This fraction was flooded with the alteration products of the hornblende, which range from colorless plates and fibers to yellowish and brownish material. The color and the general opacity of this material are largely caused by the secondary iron oxides coating the grains or in intimate intergrowths. The alteration products have a rather wide range of refractive indices. Limiting values determined on almost colorless fibers are, approximately,

[36] N. H. Darton and Sidney Paige, "Central Black Hills, South Dakota," *U.S. Geol. Surv. Folio 219* (1925), p. 4.

[37] J. J. Runner, "Intrusive Sedimentary Amphibolites" (Abst.), *Prelim. List of Titles and Abstracts, 49th Annual Meeting, Geol. Soc. Amer.* (1936), p. 42.

alpha = 1.518 and gamma = 1.550, corresponding to beidellite in the clay minerals.

TABLE 16

HEAVY-MINERAL ANALYSES

	Sample No. 13 (Fresh)	Sample No. 14 (Weathered)
	Percentage weight	
Loss on washing	14.4	30.8
Heavy minerals in washed sample	91.9	30.7
	Percentage approximate weight of total heavy minerals	
Mineral:		
Hornblende	77.0	6.7
Black opaque mineral (Ilmenite?)	14.0	18.0
Garnet	5.6	45.0
Chlorite	2.2	3.4
Leucoxene	0.72	n.d.
Epidote	0.47	0.02
Apatite	0.39	0.54
Titanite	0.03	0.03
Pyrite	0.01	0.01
Altered hornblende	.....	26.0*
Unknown	.....	0.14

* Contains leucoxene and secondary iron oxides.

According to Ross and Kerr,[38] beidellite or a related clay mineral may be formed by the alteration of ferromagnesian minerals. They state that a brownish color in thin section may indicate ferric iron or the nontronite molecule.

[38] C. S. Ross and P. F. Kerr, "The Clay Minerals and Their Identity," *Jour. Sed. Petro.*, Vol. I (1931), p. 62.

The moisture content of sample No. 14 is suggestive of the behavior of minerals in the montmorillonite-beidellite group, which characteristically adsorb water. After assigning oxides to calcite, quartz, garnet, and chlorite in amounts estimated from thin section and heavy separates, there is evidence that the hornblende has alumina enough to make beidellite a probable product.

SUMMARY

Oxidation, hydration, and solution were all active in the decomposition of the Black Hills amphibolite.

Study of the minerals in the fresh and weathered phases shows the complete removal of calcite and the severe alteration of hornblende to beidellite or related clay minerals. Garnet has resisted the alteration as is indicated by its concentration in the heavy mineral fraction of the weathered rock.

Although the *Al-C-Alk* diagram clearly brings out the chemical changes effected by weathering with the development of "sedimentary" chemical characteristics in the weathered phase, it is not certain that the diagram differentiates all para- from orthometamorphic rocks.

DISCUSSION OF PRINCIPLES

PREVIOUS WORK

In view of the great number of contributions which have swelled the literature pertaining to weathering, it is indeed fortunate that a number of compilations and treatises summarizing the processes and results of weathering are available. The works of Van Hise,[39] Steidtmann,[40] Merrill,[41] Leith and Mead,[42] Clarke,[43] Twenhofel,[44] and others are well known and readily available. The writer has attempted to correlate the findings of the present study with the conclusions reached by these men.

[39] C. R. Van Hise, "A Treatise on Metamorphism," *U.S. Geol. Surv. Mon. 47* (1904).

[40] Edward Steidtmann, "A Graphic Comparison of the Alteration of Rocks by Weathering with Their Alteration by Hot Solutions," *Econ. Geol.*, Vol. III (1908), pp. 381–410.

[41] *A Treatise on Rocks, Rock-weathering, and Soils.* [42] *Op. cit.*

[43] F. W. Clarke, "Data of Geochemistry," *U.S. Geol. Surv. Bull. 770* (1924), pp. 479–543.

[44] W. T. Twenhofel and collaborators, *Treatise on Sedimentation* (2d ed.; Baltimore: Williams & Wilkins Co., 1932), pp. 1–21.

CHEMICAL CHANGES

The rate of rock decomposition depends upon a number of variable factors. Climate and rock composition are the most important, but other conditions also require consideration in studies of individual rocks. The climatic control of weathering is emphasized by soil workers, but attempts to estimate climatic conditions of past geological time from weathering products have not met with any great degree of success. The geologist must content himself largely with the study of the changes that can be measured in the rock materials.

Students of weathering have concluded that all constituents, excepting water, are lost in prolonged chemical action. Direct com-

TABLE 17

ORDER OF THE RATES OF LOSS OF CHEMICAL CONSTITUENTS ASSUMING ALUMINA CONSTANT

	Morton Gneiss	Medford Diabase	Black Hills Amphibolite
1	Na_2O	CaO	CaO
2	CaO	MgO	Na_2O
3	MgO	K_2O	MgO
4	K_2O	Na_2O	Fe
5	SiO_2	SiO_2	SiO_2
	Al_2O_3	Al_2O_3	Al_2O_3
	Fe (gain)	Fe (gain)	K_2O (gain)
	H_2O (gain)	H_2O (gain)	H_2O (gain)

parison of chemical analyses fails to give an accurate understanding of the changes produced by weathering. Analyses of fresh and weathered material must be reduced to a common basis. The consensus of opinion is that of the common rock constituents, alumina and iron are least affected.

Steidtmann[45] concluded that the order in which the constituents are lost is lime, magnesia, soda, potash, silica, iron, and alumina. This sequence is approximately that found by other writers. The order given by Leith and Mead[46] for acid igneous rocks is lime, soda, magnesia, potash, silica, iron, and alumina.

In Table 17 the order of the rates of loss of the important constituents is given for the three rocks studied in which decomposition has been important. As is indicated by the similarity of the chemical

[45] *Op. cit.*, p. 406.

[46] *Op. cit.*, pp. 16–17.

changes noted by many investigators, the processes of weathering operate with appreciable uniformity.

STABILITY OF ROCK-FORMING MINERALS

A corollary of the concept of an order for the rates at which chemical constituents are lost in rock-weathering is that the rock-forming minerals exhibit different degrees of stability. This second concept in its broader implications is undenied. In the present study the relative stability of the original minerals has been suggested. It would seem desirable to state some order of stability applicable to the common igneous rock-forming minerals in general; and the consistency of the findings, observed and experimental, which have been reported in the literature[47] is favorable to the idea that some such order does exist.

To the list of minerals encountered in this study a few additional minerals on which numerous observations have been made can be added. It is generally agreed that olivine yields readily to weathering and that the rate of attack of this mineral exceeds that of any of the other common ferromagnesian silicates. Augite, relatively more stable than olivine, decomposes more readily than hornblende, which in turn is less stable than biotite. Potash feldspars are more resistant than the soda-lime feldspars, and sodic plagioclase is more durable than calcic plagioclase. Of the aluminum silicates comprising the essential minerals of the common igneous rocks, muscovite appears to be most resistant to decomposition.

As a group, the mafic minerals are less stable than the salic minerals. For this reason, and for others, as will become apparent, the rock minerals have been arranged in two series in order of increasing stability in Table 18. To the petrologist it is at once evident that the arrangement in the table is that of the reaction series as presented by Bowen.[48] The table, however, is not to be interpreted in the manner of the reaction series. It is not to be inferred that olivine weathers to pyroxene but rather that, in a normal igneous rock containing olivine and pyroxene, the rate of decomposition of the olivine may

[47] Good summaries are those of Clarke, *op. cit.*, pp. 481–94; and P. G. H. Boswell, *On the Mineralogy of Sedimentary Rocks* (London: Murby & Co., 1933), pp. 37–47.

[48] N. L. Bowen, "The Reaction Principle in Petrogenesis," *Jour. Geol.*, Vol. XXX (1922), pp. 177–98.

be expected to exceed that of the pyroxene. Similarly, all other things being equal, the rate of decomposition of gabbro will exceed that of granite. This concept is here called, by analogy with the reaction principle, the *stability principle*.

The coincidence between the arrangement in Table 18 and the reaction series is certainly more than fortuitous. The reaction series indicates stability of the minerals under conditions of equilibrium of composition, pressure, and temperature which prevail at the time of formation of the igneous rock-forming minerals. These conditions are markedly different from surface conditions; for this reason, min-

TABLE 18

MINERAL-STABILITY SERIES IN WEATHERING

Olivine	
	Calcic plagioclase
Augite	
	Calci-alkalic plagioclase
Hornblende	Alkali-calcic plagioclase
	Alkalic plagioclase
Biotite	
	Potash feldspar
	Muscovite
	Quartz

erals weather. Perhaps the differential between equilibrium conditions at the time of formation and those existing at the surface governs the order of stability.

The stability principle is a generalization from accumulated data. Because of the variety of conditions and agents of weathering, exceptions may be expected. The writer believes that such exceptions, rather than disproving the validity of the principle, indicate that unusual conditions attended the weathering. Careful study should be made to ascertain these conditions.

CHEMICAL CONSIDERATIONS

Chemical weathering.—The processes of chemical weathering—decomposition—are commonly stated to be solution, hydration, carbonation, and oxidation. "Hydration" is used in a general sense by geologists to cover the weathering effects of the union of water and

rock minerals. This is said to be the most important process of decomposition.

The addition of water to anhydrite to form gypsum according to the equation (1) is simple hydration:

$$(1) \qquad CaSO_4 + 2H_2O \rightarrow CaSO_4 \cdot 2H_2O \,.$$

The reaction, between feldspar and water, however, involves the decomposition of both and is not hydration in the sense of equation (1). This reaction is hydrolysis, and by soil workers it is considered the most important of the processes of decomposition.[49]

A recent paper[50] dealing with the decomposition of minerals by water summarizes the older work and presents new data in an admirable manner. By grinding silicate minerals in water and measuring the hydrogen-ion concentration resulting from the hydrolysis of the minerals, Stevens obtained results which he believes are a rough index to their weathering qualities. A number of the rock-forming minerals in Table 18 were investigated, and there is a noteworthy correlation between the results obtained by Stevens and the suggested mineral-stability series.

The role of carbon dioxide.—That carbon dioxide or carbonic acid enters into weathering processes in some significant capacity is doubted by few, and experimental data show the greater effectiveness in the attack of silicates by water containing carbon dioxide than by pure water. The question immediately arises as to how the presence of carbon dioxide would affect the process of hydrolysis, and here, indeed, the results of Stevens' work appear to be highly pertinent.

The hydrolysis of orthoclase (for example) may be represented by the equation

$$(2) \qquad 2KAlSi_3O_8 + 2HOH \rightarrow 2KOH + 2HAlSi_3O_8 \,.$$

If an aluminosilicic acid is formed according to equation (2), it would be highly unstable and decompose, yielding colloidal silica

[49] G. W. Robinson, *Soils: Their Origin, Constitution, and Classification* (London: Thomas Murby & Co., 1936), p. 44.

[50] R. E. Stevens, "Studies on the Alkalinity of Some Silicate Minerals," *U.S. Geol. Surv. Prof. Paper 185-A* (1934).

and a colloidal complex. The latter under certain conditions crystallizes, forming kaolinite, while the colloidal silica is removed by solution.

Stevens ground orthoclase in an agate mortar with one drop of phthalate buffer solution at pH = 5.0 and found that after three minutes of grinding the acidity of the buffer had been neutralized. The solution became alkaline, pH = 8.8, and, at this hydrogen-ion concentration, equilibrium was established. Grinding orthoclase with an alkaline buffer solution, pH = 10.0 resulted in no change in pH — the orthoclase was not attacked.

Returning to equation (2), supposing carbon dioxide should be bubbled though the solution, potassium hydroxide and carbonic acid would react to form carbonate and water. The result is that hydroxyl-ions are removed, and the alkalinity is thereby appreciably reduced. Reaction (2) continues until the carbonic acid is neutralized and until equilibrium is again established. The role of carbon dioxide may be chiefly in the control of alkalinity rather than in the direct attack of the carbonic acid. Other possible effects of the carbon dioxide on the solubilities of the salts and on the nature of the clay mineral produced are not considered here.

Further chemical study of the changes produced by weathering is needed, but it is particularly desirable that such work be correlated with the quantitative study of the actual mineral changes. This line of investigation has been neglected largely because of the difficulties encountered in the quantitative estimation of minerals in weathered materials and residual clays. New methods and techniques are needed to facilitate quantitative mineralogical study. Quantitative mineralogical data should lead to a better understanding of the relative stability of minerals and of the physical and chemical conditions which influence their susceptibility to alteration.

THE EFFECT OF PRESSURE ON THE RIGIDITY OF ROCKS. I

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ABSTRACT

A dynamical method has been adapted to the measurement of the velocity of torsional waves in cylinders of rock exposed to pressures as high as 4,000 kg/cm², at 30° C. and at 100° C. From these results the rigidities of the rocks under these conditions are derived, as well as approximate values for the pressure and temperature coefficients of velocity and rigidity. Very large changes of rigidity are observed in many cases upon the application of the first few hundred atmospheres; at high pressure the change of rigidity with pressure becomes nearly linear and small. The following questions are discussed: stress conditions in aggregates of crystals, the effect of anisotropy upon velocity, the calculation of other elastic parameters from rigidity and compressibility, damping and dispersion in rocks, the effect of combined pressure and temperature on velocity in the earth's crust, and the identification of materials in the crust by comparison with seismological data.

INTRODUCTION

The identification of the materials of the inaccessible depths of the earth's crust has hitherto depended upon a correlation, guided by geological theory, of the velocities found for seismic disturbances with values of compressibility of rocks determined in the laboratory, and chiefly at the Geophysical Laboratory of Washington. While the compressibility is of interest in itself in affording estimates of the change of density of definite rock types when subjected to the pressures computed for various depths in the earth, still the chief use of compressibility has been in effecting comparisons with the seismic-wave velocities. The importance of compressibility data has been due partly to the fact that it has not been possible until recently to measure any other elastic coefficient of rocks under high pressure. For the purpose of finding the velocities in given rock samples under the conditions existing in the upper part of the crust, the compressibility data are incomplete in three respects. First, the velocities are not determined by the compressibility alone; it is necessary to know one other elastic parameter. This difficulty has been met by assum-

[1] Paper No. 38, published under the auspices of the Committee on Geophysical Research and the Division of Geological Sciences at Harvard University.

ing a value for Poisson's ratio σ, which apparently does not vary enormously in the crust, and which is obtained from seismological results. Second, where the compressibility is small, it has not been possible to measure accurately its change with pressure. Third, the effect of temperature upon compressibility is not known for rocks.

The present paper describes direct measurements of the velocity of torsional waves in rocks under moderately high pressure; this velocity may be compared immediately with the velocities of the *S* or equivoluminal waves of seismology. The modulus of rigidity may be derived from this velocity, and thus, if the compressibility is known, the elastic behavior is completely specified wherever two parameters are sufficient. The sensitivity of this method is great enough to afford rough values for the change of rigidity with pressure; the effect of temperature from 30° to 100° C. at 4,000 atmospheres has also been obtained for a few samples of each of the more important rock types.

A number of investigators have studied the behavior of rocks at atmospheric pressure, using the methods of testing common in engineering practice, such as bending, twisting, and compressing under static loads, bending and twisting periodically by impact or otherwise, bouncing spheres from a polished surface, etc. The results of these tests, while generally in qualitative agreement, show very great numerical discrepancies for rocks supposed to be similar. In some cases values of elastic parameters have been obtained which were mutually inconsistent in terms of the theory of an elastic isotropic body, which is used in such cases mainly for purposes of reference. Probably the most useful results using ordinary statical methods are the relatively old ones of F. D. Adams and E. G. Coker[2] and the recent ones of W. A. Zisman.[3] These writers measured Young's modulus and Poisson's ratio by finding the changes of dimensions of columns subjected to axial compression; Zisman also measured the compressibility in order to study the relations between the elastic param-

[2] "An Investigation into the Elastic Constants of Rocks, More Especially with Reference to Cubic Compressibility," *Carnegie Inst. Washington Pub. 46* (1906).

[3] "Young's Modulus and Poisson's Ratio with Reference to Geophysical Applications," "Compressibility and Anisotropy of Rocks at and near the Earth's Surface," "Comparison of the Statically and Seismologically Determined Elastic Constants of Rocks," *Proc. Nat. Acad. Sci.*, Vol. XIX (1933), p. 653.

eters. Zisman observed that Hooke's law usually failed to hold for the stress range (0–200 kg/cm²) which he employed in axial compression: the stress-strain relation was not linear, but the stress increased more rapidly than the strain. Young's modulus, defined as the slope of the stress-strain curve, increased with the mean stress. The measurements of Adams and Coker, on the other hand, dealing with a higher stress range (60–600 kg/cm²), showed a nearly linear stress-strain relation in most cases. This behavior was attributed to the looseness of structure of the rocks, generally referred to as "porosity." Increasing stress may be supposed to reduce this looseness, pressing the crystals into closer contact, and at some stress the initial looseness may be largely overcome, with the result that for additional stresses Hooke's law will be closely followed. This compact structure, for which the theory of the elastic isotropic solid may be considered valid, at least as a first approximation, is naturally expected to exist at the depths in the crust which are of major geophysical interest.

Comparatively few of the rocks which are collected at the surface for laboratory study are so compact as to satisfy Hooke's law. Even fresh rocks have experienced changes of pressure or of temperature which, acting unequally upon the different kinds of crystals, have loosened the structure. Preparation of samples for the laboratory may produce further deterioration, and many forms of testing cause more or less permanent changes. For these reasons the usual methods of testing have generally failed to give results of much value for geophysics. The most pertinent as well as the most complete and far-reaching results have been the measurements of compressibility at very high pressures by L. H. Adams and his collaborators, for which complete references will be given later. The technique of this work was such that the structure of the rock played a relatively unimportant part; as we have seen, the supplementary determination of another elastic parameter, under comparable conditions, is required for completeness.

Measurement of the velocity of propagation of elastic waves in metals and glasses has long been a familiar experiment in general physics. Since the days of Kundt's tube, the technique of such measurements has been very greatly improved, and recent results, ob-

tained with modern methods, have been published by G. W. Pierce, D. S. Muzzey, S. L. Quimby, J. M. Ide, and others. Muzzey[4] and Ide[5] have applied these new methods to the study of wave velocities in a number of rocks, chiefly to those specimens which were studied with static methods by Zisman. Ide found the velocities of both longitudinal and torsional waves and has published a detailed comparison of these results with those found by Zisman. As before, it was found that the elastic parameters derived from these velocities failed to satisfy the equations of elasticity, except for one unusually compact rock (Vinal Haven diabase). It appeared probable, however, that this technique could be applied to rocks when they were subjected to conditions which would ensure their compactness, that is, under high pressure.

With this object Dr. Ide and one of the present writers attempted for a considerable period to find the effect of pressure on the frequency of longitudinal vibrations of rods. While it was possible to produce such vibrations and to observe their apparent resonant frequencies under pressures as high as 8,000 atmospheres and presumably much higher, it was found that the interaction of the pressure medium with the vibrating rods varied periodically as the pressure was increased, in such a way as to introduce very large uncertainties as to the proper frequency of the rod alone. This phenomenon, the interference of a sound field with a resonator, is well known in ordinary acoustics, that is, with air at 1 atmosphere as the pressure medium. With a pressure medium at several thousand atmospheres, the effect upon the apparent resonant frequency of the source may be extremely large, and it is difficult, if not impossible, to calculate this effect with satisfactory precision. In consequence, the study of longitudinal vibrations under pressure was abandoned, at least temporarily, in favor of the study of torsional vibrations.

The torsional vibrations of a circular rod involve only tangential motions of the surface; there is no volume change and no motion normal to the surface. The interaction of the vibrating rod with the

[4] Quoted by W. A. Zisman, *ibid.*

[5] Ide, "Comparison of Statically and Dynamically Determined Young's Modulus of Rocks," *Proc. Nat. Acad. Sci.*, Vol. XXII (1936), pp. 81–92; "The Elastic Properties of Rocks: A Correlation of Theory and Experiment," *ibid.*, pp. 482–96.

pressure medium depends therefore only upon the viscous forces acting at the surface of the rod. The shear wave sent out into the medium is very rapidly damped out, so that only a very thin skin of the medium is active in perturbing the motion of the rod; its effect upon the frequency, which can be computed with sufficient precision, turns out to be of a negligible order of magnitude. As a preliminary to the study of rocks, a study was made of the torsional vibrations of several metals and glasses under pressure; these results, together with a description of the method, have already been published.[6] For the present work with rocks the same apparatus was employed, except that more powerful driving and receiving units were required. In order to shorten an already long publication, no further detailed description of the apparatus will be included in this paper.

The principle of these measurements may be very briefly reviewed. The specimen, a right circular cylinder, usually about 20 cm long and 2 cm or less in diameter, is set into forced torsional oscillation by means of an alternating torque applied to a small driving armature at one end. The resulting motion is observed with the aid of another armature at the other end, moving in the field of coils mounted on the poles of a permanent magnet, and connected to an amplifier, whose output is indicated by an A.C. voltmeter. The known frequency of the applied torque is varied until resonance of the vibrating system is indicated by the large deflection of the voltmeter. From the resonant frequency of the rod and its length, the velocity is readily computed by the well-known relation, $v = 2fL$, if f is the lowest resonant frequency or fundamental. In practice, small corrections must be applied to the observed resonant frequency to take account of the loading due to the armatures and other perturbations. These corrections will be dealt with in a later section. A rough measure of the internal damping in the specimen may be derived from the shape of the resonance curve.

The vibrating system and driving and observing units are mounted in a thin brass tube and placed in a heavy-walled steel cylinder, in which they may be subjected to the pressure of nitrogen generated

[6] Francis Birch, "The Effect of Pressure on the Modulus of Rigidity of Several Metals and Glasses," *Jour. App. Phys.*, Vol. VIII (1937), p. 129.

by a suitable hydraulic press of the type employed by Bridgman,[7] to whose work all questions concerning pressure technique may be referred. For convenience, the pressure range of the present work is, from 1 to 4,000 kg/cm². There is no fundamental reason, however, which prevents the extension of these measurements to much higher pressure. Unless otherwise indicated, all work was done at 30° C.; this temperature was maintained by circulating a stream of water from a thermostatically controlled supply through a water jacket around the pressure cylinder. In some cases the temperature was changed to nearly 100° C. by circulating boiling water through the jacket.

STRESS CONDITIONS; TREATMENT OF SPECIMENS

Previous work[8] with rocks under pressure has shown that very different effects may be produced, depending upon whether the surface of the rock is exposed to the pressure medium or is protected from contact with the medium by means of a flexible, impermeable covering. It is not difficult to see why this must be so. If an impermeable covering is used, or if the rock itself is completely impermeable, the stress is hydrostatic over the external surface of the rock; the rock is, however, a porous aggregate of crystals, usually of different kinds, and the stress on any individual crystal will not in general be hydrostatic. On the contrary, the stress will at first be concentrated in those regions which are in solid contact. Increasing pressure will increase the areas of contact between crystals until equilibrium is reached, but, except in the special case that the crystals are all alike and belong to the cubic system, the stress on the individual crystals will not in general correspond to hydrostatic pressure. Under these conditions one would expect to find evidence of hysteresis and permanent set, and this is indeed observed.

Now if the rock is exposed to the pressure medium, the medium will in nearly all cases penetrate the intercrystalline spaces or "pore" spaces at high pressure. A fairly definite pressure is usually required

[7] P. W. Bridgman, *The Physics of High Pressure* (New York: Macmillan, 1931).

[8] Th. von Karman, "Festigkeitsversuche unter allseitigen Druck," *Zeit. d. Ver. deutscher Ingenieure*, Vol. LV (1911), pp. 1749–57; L. H. Adams and E. G. Williamson, "The Compressibility of Minerals and Rocks at High Pressures," *Jour. Franklin Inst.*, Vol. CXCV (1923), pp. 475–529. W. A. Zisman, *op. cit.*

before penetration begins with a rock which has not been exposed to such pressure before. But once penetration begins, the stress on the crystals changes its character and begins to approach a hydrostatic pressure. The more closely this condition is realized, the less are the crystals forced into mutual contact by the pressure; the rock becomes an open network of crystals, each under nearly hydrostatic pressure and consequently shrinking away from its neighbors. Under these conditions the porosity and permeability may increase with the pressure.

It is clear that the condition which one hopes to realize by high-pressure experimenting is that produced by transmitting pressure through an impermeable covering, since in this case only will the pressure have the effect of closing up the structure of the rock and of improving its mechanical properties. The other condition, that of high hydrostatic pressure exerted upon the individual crystals by a permeating fluid, may be realized under special conditions in nature, but it does not appear to resemble the conditions at great or even moderate depths in the crust. The conditions are of course indistinguishable for metals or glasses, and occasionally a rock specimen has been found to be very nearly impermeable, at least until it has been under pressure several times. Perfect permeability and perfect impermeability are merely the two extreme possibilities, of which only the latter can be realized at will, experimentally, by the use of a covering. With uncovered rocks nearly every degree of permeability may be encountered, and for a given rock the permeability will change with every change of conditions.

It is at least possible that in any impermeable aggregate of any kinds or proportions of crystals in equilibrium with a particular hydrostatic pressure at some boundary surface and at constant, uniform temperature, the arrangement and sizes of the crystals may be such that the stress in every single crystal is also the same hydrostatic pressure. We can see this by considering the aggregate first divided up into its component crystals and all the parts subjected to the given pressure by means of a fluid. Then each crystal will come to equilibrium with the pressure, assuming a definite size and shape. We may now imagine a Maxwell demon stonemason who will fit the pieces together under pressure in such a way as to form a perfectly

compact aggregate with no spaces, and hence no fluid, between the crystals. He can do this because he can cut any of the crystals to any shape necessary without changing the stress. The result is finally an aggregate from which the pressure medium is excluded; not only is there equilibrium with the hydrostatic pressure over its boundary, but each crystal is also in equilibrium with the same hydrostatic pressure. The aggregate may now be enclosed in an impermeable coating without causing any change of stress.

But now suppose that the pressure over the boundary is changed, reduced, for example. Then the more compressible crystals expand more than the less compressible ones, and we have two alternatives, depending upon the boundary conditions: if the boundary remains impermeable (or if the pressure is exerted by solid contact), the aggregate may remain nearly compact, but the stresses on the crystals are no longer hydrostatic; if the pressure is exerted by a fluid which can penetrate the aggregate, the stresses on the crystals can remain hydrostatic, but the aggregate will no longer be compact—spaces must open up between the crystals. If the pressure is increased, corresponding results follow. In other words, for a given impermeable aggregate with definitely fixed crystals, there exists in general only one pressure (for a given temperature) at which the stresses on the individual crystals can be hydrostatic and equal to the pressure on the boundary, and it is of course possible that there will not be even one such pressure. At all other pressures the stresses on the crystals must depart more or less from hydrostatic.

This is true of homogeneous as well as heterogeneous aggregates, except where the crystals belong to the cubic system. It can be demonstrated with polycrystalline zinc, for example;[9] here the changes of length of a zinc rod under pressure show distinct hysteresis effects, attributable to non-hydrostatic stress upon the individual crystals.

If the conditions are such that plastic flow or recrystallization are possible, then there will be a tendency to return to hydrostatic pressure in the impermeable compact aggregate after a change of pressure (or of temperature, which will produce similar effects). Very probably such crystalline aggregates as have existed for long periods

[9] P. W. Bridgman, "The Compressibility of Thirty Metals as a Function of Pressure and Temperature," *Proc. Amer. Acad. Arts and Sci.*, Vol. LVIII (1923), p. 166.

at considerable depths in the earth's crust have come close to perfect hydrostatic equilibrium. This is a special case of the process of annealing, which may in general be considered as a process of removing non-hydrostatic strains.

Certain examples of the partial annealing of rocks at high pressure due to heating to the relatively low temperature of 100° C. will be cited in the section dealing with the effect of temperature. It will also appear that the beneficial effects of such annealing do not necessarily remain when the pressure is altered; this is a consequence of our general theorem, which would require a new annealing at each different pressure.

One of the most curious effects observed with the uncovered rocks is an abnormal expansion of the rock when the external pressure is reduced; a certain time is required for the pressure to equalize itself inside and outside the rock, and, during this time, the aggregate is exposed to an internal pressure greater than the external pressure. This is likely to produce permanent elongation at low pressure, and reduction of the wave velocities. A specimen of Sudbury norite exposed to high nitrogen pressure was placed, after fairly rapid reduction of the pressure, in a test tube full of water; enough nitrogen was left in the rock to produce a profusion of bubbles for perhaps half an hour.

We are now able to give an explanation of the low values of compressibility found for olivine diabase from Vinal Haven, Maine, by Birch and Dow.[10] These values were obtained by observing the change of length of a column of the material exposed to the pressure medium. Under these conditions the overall change of length of the rod was determined by the compressibility of the least compressible elements and by the crystals whose direction of lowest compressibility lay parallel to the axis of the rod. The elements of higher compressibility were freely compressed by the permeating medium, but the observed change of length depended only upon the skeleton of the less compressible crystals. For this reason, although the observations were correct and have been repeated with numerous other

[10] Francis Birch and R. B. Dow, "Compressibility of Rocks and Glasses at High Temperatures and Pressures: Seismological Application," *Bull. Geol. Soc. Amer.* 47 (1936), p. 1235.

specimens, and although the material was almost perfectly isotropic, the values obtained do not represent the mean compressibility of all the elements, and hence are not representative of the rock. The values of velocity derived from these compressibilities are therefore incorrect. This whole question will be treated in more detail in another paper.

In addition to the stresses due to the hydrostatic pressure there are the stresses involved in the torsional motion of the rod. These are relatively small and are incapable of producing perceptible permanent changes. The strains due to this motion are probably not greater than 10^{-6} and may be appreciably smaller; thus they are not greatly different in magnitude from the strains involved in the transmission of seismic waves, at least at a considerable distance from the source.

The realization of a light flexible impermeable covering for the conditions of these experiments is a matter of some difficulty. Copper foil and tin foil have been used by former workers. Hoping to avoid some of the labor involved with these, the present writers tried out such materials as picein wax, glyptal cement, aluminum paint, and a kind of rubber paint. None of these was sufficiently impermeable to the compressed nitrogen. It is very easy during a run to tell whether the gas is penetrating the rock or not. If there is no penetration the resonant frequency increases as the pressure increases, and after each change of pressure there is a period during which a small amount of after-working takes place and the frequency increases a little more. But if a change of pressure induces a leak in the covering, penetration begins and the frequency commences to decrease; the amount of the decrease may be very large. It is thus possible to know very soon if the covering is tight.

The copper-foil covering was finally adopted. This consists of a sheet of 0.002-inch copper foil wrapped around the rock, soldered along one side, and soldered at the ends to disks or spun caps of copper. The soldering must be done with extreme care, and the foil must be examined for occasional pinholes. It is, furthermore, a great help in securing a good proportion of successful covers to fill the fine irregularities of the surface of the specimens with plaster of Paris, especially in the case of the looser, coarse-grained rocks; any excess

is scraped off after the plaster sets, leaving a smooth surface for the copper foil to bear upon. A high polish on the surface of the specimens might have been advantageous; in general the specimens were cored from massive blocks of the material and reduced to uniform circular cylinders in the lathe, with smooth but not polished surfaces. Without the plaster of Paris there was danger that the foil would be forced too far into small surface cracks and be ruptured. This often happened even when every precaution had been taken.

The driving armatures and mounting pins are soldered to the copper end-pieces. The whole copper sheath fits the rock very closely, especially after a little pressure has been applied; this method of attaching the armatures is almost the only way of attaching them securely to a rock for use at high pressure. It has also proved most satisfactory with glasses, where other methods have proved unreliable.

CORRECTIONS AND CALCULATIONS

The question of the corrections required for the effect of the armatures and mounting pins upon the resonant frequency of the vibrating system has already been treated.[11] When these corrections are small, it is sufficient to find the contribution to the kinetic and potential energies of the system due to these parts, on the assumption that the form of the vibration is unchanged by their presence. The correction for the copper sheath may be found in the same way. It is also possible in a number of cases to obtain the frequency with and without the sheath, and to compare the difference thus observed with the calculated correction. The calculated correction is based on the assumption that the copper foil follows without slip the motion of the surface of the rock and that this motion is of the same form as for the free vibration of a homogeneous elastic rod. These assumptions are justified by the approximate agreement of the observed and calculated corrections.

The kinetic and potential energies of the rod may be expressed in terms of a single variable ϕ, the angular displacement at either end of the rod. For the rod alone, we have the kinetic energy $T = a\dot{\phi}^2/2$, and the potential energy $U = c\phi^2/2$, with $a = MR^2/4$ and $c = \pi^3\mu R^4/4L$, where L is the length, R the radius, M the mass, and μ

[11] Birch, *op. cit.*

the rigidity of the rod. The resonant frequency f_0 is then given by $2\pi f_0 = (c/a)^{1/2}$. In the case of small additional masses and constraints, the actual frequency f will be given approximately by $f^2 = f_0^2(1 + \Delta c/c)/(1 + \Delta a/a)$, where Δa and Δc are the contributions to the coefficients of the kinetic and potential energies, respectively, due to these masses and constraints. This relation may usually be reduced to the form $f_0 = f[1 + (1/2)(\Delta a/a - \Delta c/c)]$.

The foil on the sides of the rod adds a fraction $2m/M$ to $\Delta a/a$, and $4\mu'\Delta R/\mu R$ to $\Delta c/c$, where μ' is the rigidity, m the mass, and ΔR the thickness of the foil on the sides (in which is included the mass of the solder along the sides). For the parts at the ends, the contribution to $\Delta a/a$ is $4m'k^2/MR^2$, where m' is the mass and k the radius of gyration of these parts about the axis. It is hard to compute k exactly; it lies between R and $R/\sqrt{2}$, depending upon the distribution of solder over the ends and around the rim. Similarly, the line of solder along the side introduces some uncertainty in $\Delta c/c$. The uncertainty in the absolute value of the frequency of the rod alone is about $\frac{1}{2}$–1 per cent from these corrections, whose total amount is not more than 7 per cent. The frequency calibration of the oscillator is known to about $\frac{1}{2}$ per cent over the entire range, so that in the end the frequency of the rock alone is known with an uncertainty of not greater than about 1.5 per cent. The kinetic energy correction does not change appreciably with pressure; the potential energy correction, which involves the rigidity of the rock, may change considerably at the lower pressures when the rigidity change in this region is large. An experimental check of the effect of pressure on the corrections was carried out by using a covered aluminum specimen. This was the same specimen previously used without cover,[12] and the pressure coefficient of rigidity was found to be the same as before.

If then f_0 is the frequency of the rock alone, the velocity of the torsional wave is given by $v = 2f_0L$. In order to find the variation of velocity, the variation of length as well as of frequency should be taken into account. The changes of length of most of these rocks under these conditions have been measured and will be published separately. The change of length becomes relatively important only

[12] *Ibid.*

if we wish to form the pressure coefficients of velocity or of rigidity. These are given by

$$\frac{1}{v}\left(\frac{\Delta v}{\Delta p}\right)_t = \frac{1}{f_0}\left(\frac{\Delta f_0}{\Delta p}\right)_t + \frac{1}{L}\left(\frac{\Delta L}{\Delta p}\right)_t$$

and

$$\frac{1}{\mu}\left(\frac{\Delta \mu}{\Delta p}\right)_t = \frac{2}{f_0}\left(\frac{\Delta f_0}{\Delta P}\right)_t - \frac{1}{L}\left(\frac{\Delta L}{\Delta p}\right)_t .$$

If we wish only the absolute velocity at any pressure, the change of length is unimportant in comparison with the uncertainty in the absolute frequency, since the relative change of length is only a few tenths of 1 per cent up to 4,000 atmospheres. The pressure coefficients, however, may be obtained with greater precision than this error in the absolute frequency would indicate, since they do not depend upon the oscillator calibration, nor to a large extent upon the jacket correction. At the highest pressures, however, the changes of frequency for 1,000 atmospheres become of the same order as the relative precision of frequency measurement, which is never better than about 1 part in 6,000. The values for

$$\frac{1}{v}\left(\frac{\Delta v}{\Delta p}\right)_t$$

are therefore not sure to be better than about 30 per cent at the highest pressure, in some cases.

DESCRIPTION OF MATERIALS

The rocks which have been studied are listed, with their places of origin, in Table 1; the chemical analyses are given in Table 2; and the mineral compositions, determined from thin sections, in Table 3. In Table 1 are also given the densities obtained from the weight and dimensions of carefully worked specimens.

Petrographic descriptions of the Vermont marble, quartzitic sandstone, Rockport and Quincy granites, Pelham gneiss, Sudbury norite, Vinal Haven diabase, French Creek "norite," and Balsam Gap dunite may be found in the papers of Zisman, and are not repro-

duced except for the summaries in Table 3. For these rocks we used some of the same samples which were used by Zisman for compressibility determinations, with new ones to clear up certain questions. The new samples of Vinal Haven diabase were not made from the same block but cannot be very different. For the Maryland diabase, which comes from the same locality from which the sample used by

TABLE 1

MATERIALS STUDIED, PLACES OF ORIGIN, AND DENSITIES

Material	Location	Density
Limestone	Solenhofen, Bavaria	2.605
Marble	Proctor, Vt.	2.71
Quartzitic sandstone	Allentown, Pa.	2.66
Granite	Quincy, Mass.	2.64
Granite	Rockport, Mass.	2.62
Granite	Westerly, R.I.	2.64
Gneiss	Pelham, Mass.	2.64
Syenite	Peninsula Station, Ont.	2.79
Norite	Sudbury, Ont.	2.86
Diabase	Vinal Haven, Me.	2.962
Diabase	Frederick, Md.	3.013
"Norite" (gabbro)	French Creek, Pa.	3.033
Gabbro	Mellen, Wis.	2.90
Pyroxenite (hypersthenite)	Stillwater Complex, Mont.	3.272
Pyroxenite (bronzitite)	Pilandsberg, Transvaal	3.289
Dunite	Balsam Gap, N.C.	3.275
Obsidian	Modoc, Calif.	2.440
Cores from the Noranda Mines, Ltd., Ontario		
Diabase (Keweenawan)?	2,500 feet from surface 5	2.989
Metadiabase	1,750 feet from surface 3	3.039
Metadiabase	1,750 feet from surface 7	2.989

Adams and Gibson[13] was obtained, we are greatly indebted to Dr. R. W. Goranson of the Geophysical Laboratory, and we have taken the analysis for this material from the paper of Adams and Gibson. New descriptions, for which the writers are indebted to Dr. C. S. Hurlbut, of the Department of Mineralogy, Harvard University, are given for the Ontario syenite, Transvaal pyroxenite, Stillwater pyroxenite, Mellen gabbro, and Westerly granite. New chemical analyses, kindly carried out for this paper by Dr. F. A. Gonyer, are given

[13] L. H. Adams and R. E. Gibson, "The Elastic Properties of Certain Basic Rocks and of Their Constituent Minerals," *Proc. Nat. Acad. Sci.*, Vol. XV (1929), p. 713.

for the French Creek "norite," Ontario syenite, Vinal Haven diabase, and Mellen gabbro. In connection with the French Creek "norite," it was the opinion of Professor E. S. Larsen, on examining the chemical analysis, that this rock should be designated diabasic

TABLE 2

CHEMICAL ANALYSES OF THE ROCKS

	Granite Quincy*	Granite Westerly†	Gneiss Pelham‡	Syenite Ontario§	Diabase Vinal Haven§	Diabase Maryland‖	Gabbro French Creek§	Gabbro Mellen, Wis.§	Pyroxenite Transvaal¶	Olivine of Balsam Gap Dunite**
SiO_2	74.86	72.26	72.45	55.80	48.24	51.28	52.80	47.62	55.30	38.36
Al_2O_3	11.61	13.58	13.32	14.68	18.12	15.07	13.04	22.48	2.20	0.58
Fe_2O_3	2.29	2.97	1.93	2.58	1.25	1.12	0.31	none	0.45	1.58
FeO	1.25	0.75	0.63	8.96	6.70	9.31	8.74	8.74	10.05	6.62
MgO	0.05	0.03	0.44	0.98	9.41	7.97	9.98	7.38	29.80	51.50
CaO	0.41	1.24	1.81	4.58	11.84	11.42	11.94	9.66	1.80	0.10
Na_2O	4.30	2.18	3.55	4.55	2.56	2.03	1.61	2.84	0.30	.....
K_2O	4.64	5.69	3.86	4.78	0.26	0.27	0.48	0.32	.....	.....
H_2O+	0.31	0.57	1.51	0.83	0.66	0.39	0.31	0.52	.....	.....
H_2O-	0.04	0.09	0.59	0.22	.....	0.09	.....	0.04	0.10	.....
CO_2	.....	.....	none	0.62	.....	.....	.....	.....	.....	.....
TiO_2	0.20	.....	0.27	0.66	0.82	0.78	0.84	0.42	0.10	.....
S	.....	.....	.....	0.15	0.12	0.06	0.03	0.04	.....	.....
NiO	.....	.....	.....	.....	.....	.....	.....	.....	0.20	0.34
MnO	0.02	.....	.....	0.12	0.08	0.16	0.12	0.06	0.15	0.09
P_2O_5	tr	.....	0.06	0.30	.....	0.13	0.07	0.09	0.20	.....
Total	99.98	99.36	100.42	99.81	100.10	100.08	100.27	100.21	100.65	99.17

* C. H. Warren, *Proc. Amer. Acad.*, Vol. XLIX (1913), p. 227

† I. A. Williams, *Amer. Geol. 35, 37* (1905), or Washington's *Tables*. This analysis is probably not applicable to our sample.

‡ G. Steiger, I. N. Dale, *U.S. Geol. Surv. Bull 470, 261* (1911), or Washington's *Tables*, p. 169.

§ F. A. Gonyer, new, unpublished analyses.

‖ E. S. Shepherd, in L. H. Adams and R. E. Gibson, *Proc. Nat. Acad. Sci.*, Vol. XV (1929), p. 713.

¶ H. G. Weall, in *The Bushveld Igneous Complex of the Central Transvaal* (Pretoria: A. L. Hall, 1932), p. 135 (Analysis III).

** F. A. Gonyer, in Ross, Shannon, and Gonyer, *Econ. Geol.*, Vol. XXIII, No. 2 (1928), p. 545.

hypersthene gabbro. Since the name "norite" has already been used for it in the papers of Ide and of Zisman, it will be retained with quotation marks.

Specimens of *Westerly granite* have been described in papers by Adams and Williamson,[14] and by Leet and Ewing.[15] Our specimen

[14] *Op. cit.*

[15] L. Don Leet and W. Maurice Ewing, "Velocity of Elastic Waves in Granite," *Physics*, Vol. II (1932), p. 160.

TABLE 3

MINERAL COMPOSITION OF THE ROCKS; VOLUME PERCENTAGES

	Quartzitic sandstone	Granite Quincy	Granite Rockport	Granite Westerly	Syenite Ontario	Norite Sudbury 2	Diabase Vinal Haven (norm)	Diabase Maryland (norm)	Gabbro French Creek	Gabbro Mellen	Pyroxenite Stillwater (Hypersthenite)	Pyroxenite Transvaal (Bronzitite)	Dunite Balsam Gap
Quartz	80	26	28	19.3		8							
Potash feldspar	1			33.1	84.3	1							
Albite		13		40.3									
Microperthite	5	50	64										
Plagioclase						67 $Ab_{48}An_{52}$	61.2 $Ab_{30}An_{70}$	48 Ab_1An_2	51 $Ab_{35}An_{65}$	72 $Ab_{40}An_{60}$		3.8 $Ab_{40}An_{60}$	
Biotite				5.9	0.4	6	3		0.5	1.2			
Amphibole*		(r)7 (s)3	(r)3 (s)3			(h)5					(h)4	(h) 1.7	
Pyroxene†					(aa) 12.9	(a)5 (hy)8	(di) 17.3 18.1	(a)24 (hy)25	(a)32 (hy)15	(d) 14.1 11.4	(hy) 93.5 1.3	(b) 92.3	
Olivine					1.6								97
Magnetite	7			0.9	0.3		3.4	3	0.5	1.1	1.2	2.2	
Rest	7 (sericite)	1	2	0.1	0.5				1	0.2			3

* Amphibole: (r) = riebeckite, (s) = soda-amphibole, (h) = hornblende.

† Pyroxene: (aa) = aegirite-augite, (a) = augite, (b) = bronzite, (d) = diallage, (di) = diopside, (hy) = hypersthene.

appeared to differ from these descriptions, so a special study was made of our sample; the chief difference appears to be in a distinctly lower quartz content, as shown in Table 3. Leet and Ewing, and also Adams and Williamson, used specimens having about 35 per cent quartz, while ours had only 20 per cent quartz, the difference being made up by an increased albite content. The chemical analysis of Table 2 does not refer to our sample. The complete description is as follows: This is a granular rock with maximum diameter of grain at 2.0 mm. and average at about 0.5 mm. The albite grains tend more than any other of the minerals to be tabular and elongated in outline. A slight zoning is present and the central portions are usually partly altered to sericite and carbonate. Individual grains of sericite thus found range up to 0.4 mm. in maximum dimension. Both microcline and orthoclase are present, with microcline in somewhat greater amount. The quartz is present in roughly equidimensional grains and does not appear to be interstitial to the feldspar. The biotite is in places altered to chlorite. Sphene, apatite, and zircon together make up about 0.1 per cent. Most of the zircon is in the biotite.

The *augite syenite*, from Peninsula Railway Station, Canadian Pacific Railway, was kindly supplied by the Cold Spring Granite Company, at the request of Professor G. A. Thiel of the University of Minnesota. It is a coarse-grained rock, mainly composed of perthitic feldspar and aegirite-augite. The lengths of the crystals range from 1 to 5 mm. and from 0.5 to 5 mm., respectively. The feldspar is chiefly an intergrowth of orthoclase and albite, but some grains are composed entirely of orthoclase while others have a rim of perthite around cores of orthoclase. The aegirite-augite has a mean index of refraction of 1.783, thus indicating a member of the soda-pyroxene group. Some grains have green borders, where there is an increase of the acmite molecule. The olivine is iron-rich; it is slightly serpentinized. The biotite is a reaction product around the augite and magnetite. A few grains of carbonate are products of alteration of the pyroxene.

The *gabbro* from Mellen, Wisconsin, was also supplied by the Cold Spring Granite Company. It is a coarse-grained rock, chiefly composed of labradorite, diallage, and olivine. The lengths of grains

range, respectively, from 1 to 10 mm., from 1 to 4 mm., and around 1 mm. The labradorite has the composition $Ab_{40}An_{60}$. Some of its crystals are more than 10 mm. long. The mineral is quite fresh, uniform in composition, with no zoning. The diallage forms angular grains interstitial with respect to the feldspar. There is a slight amount of alteration to biotite. The olivine is usually fresh; in a few places slightly serpentinized. It has kelyphitic borders against the feldspar. The biotite occurs chiefly as rims around magnetite grains.

The *pyroxenite* (*bronzitite*), from the Bushveld Complex of the Transvaal, is a granular rock made up chiefly of one mineral, bronzite, with individual grains ranging up to 4 mm. in maximum diameter, but averaging about 2.0 mm. The bronzite occurs in nearly equidimensional grains with a slight elongation parallel to the *c*-axis. For the most part it is fresh, but some slight alteration to hornblende is present. Magnetite is found in rounded grains with the average diameter about 0.2 mm. The feldspar is a labradorite with the approximate composition, $Ab_{40}An_{60}$. It occurs as irregular grains between the earlier bronzite crystals.

The *pyroxenite* (*hypersthenite*) from the Stillwater Igneous Complex was presented by Professor E. S. Larsen, who collected it. This is an essentially monomineralic rock composed of hypersthene, with maximum grain diameter of 6.0 mm. and average of 2.5 mm. The mean index of refraction is about 1.678, indicating a ratio of 6:1 for Mg:Fe. Most of the grains are large and tabular, but in places groups of smaller granular crystals are found. There is some slight alteration to hornblende, usually along cleavage planes. Exceedingly small particles of magnetite are scattered through the rock. One large grain of olivine was found in the thin section.

The Sudbury norite cores were not all made from a single block having a volume of 1 or 2 cubic feet, as in the case of the other specimens. They are random drill cores from the large body of norite, and, as shown by the analyses given in Zisman's paper,[16] they are not all alike. Chemical analyses are given in a paper by Coleman, Moore, and Walker,[17] but are not reproduced. The analysis given

[16] *Op. cit.*

[17] A. P. Coleman, E. S. Moore, and T. L. Walker, *Univ. Toronto Studies*, No. 28 (1929), p. 18.

in Table 3 refers to core No. 2 (the same numbering as in Zisman's paper has been retained where the same samples have been used).

In the case of the Vinal Haven diabase, the norm, calculated from the chemical analysis by Professor Larsen, differed perceptibly from the thin section analysis previously published,[18] especially with respect to the proportions of feldspar and of olivine. The density and mechanical properties of this rock suggest that the norm is probably more nearly correct, and this is given in Table 3.

For the "Noranda" cores we are indebted to Dr. H. M. Butterfield, formerly chief geologist of the Noranda Mines, Limited. These samples were made from drill cores taken at much greater depths than were our other specimens. Thus the "metadiabase" came from a depth of 1,700 feet, and the Keweenawan diabase from 2,500 feet. These designations were given by the geologists connected with the mines. We have no chemical or mineral analyses of these materials.

NUMERICAL RESULTS

The experimental results are expressed in the Tables 4–9. These require a little explanation, and certain peculiarities of the individual specimens should be pointed out. Table 4 contains the bulk of the data. Here the torsional velocity is given at various pressures for all of the rocks studied. These velocities are obtained by reading off the oscillator settings from a smooth curve drawn through the observed points at the desired pressure intervals; the observed points were taken at 500 atmosphere intervals at high pressure, and at smaller intervals at low pressure. These readings are reduced to frequencies with the aid of the oscillator-calibration curve or formula; the frequencies are corrected as described in a preceding section, and the velocities calculated. No correction for change of length has been made in Table 4, since this is smaller than the probable error of frequency calibration. On the other hand, this correction has been applied to the coefficients of Table 6. The measurements of velocity are absolute measurements; they do not depend on the velocity of some standard reference material, nor does any one velocity, for a given sample, depend upon the velocity at any other pressure.

In nearly all cases in the region above 1,000 atmospheres the

[18] Zisman, *op. cit.*

TABLE 4

VELOCITIES OF TORSIONAL WAVES AS FUNCTION OF PRESSURE, AT 30° C

MATERIAL	VELOCITY IN KM/SEC							
	Pressure, kg/cm²							
	1	100	200	500	1,000	2,000	3,000	4,000
Limestone, Solenhofen	2.75	.....	.....	2.91	2.97	3.04	3.06	3.08
Marble, Vermont 3	2.4	3.24	3.33	3.42	3.46	3.48	3.48	3.49
Quartzitic sandstone 2(E)	3.39	.....	.....	3.87	3.99	4.03	4.05	4.07
3	3.43	.....	.....	3.79	3.86	3.91	3.93	.....
Granite:								
Rockport (⊥)	2.54	.....	.....	3.44	3.53	3.58	3.61	3.62
(E)	2.57	3.02	3.21	3.40	3.49	3.54	3.55	3.56
Quincy (100 ft.), 1	2.77	3.19	3.31	3.47	3.56	3.59	3.61	3.62
Same, new jacket	2.81	.....	.....	3.51	3.56	3.62	3.64	3.65
(235 ft.) 4 (H)	2.52	3.15	3.28	3.45	3.54	3.57	3.60	3.62
(235 ft.) 5 (E)	2.54	3.16	3.27	3.44	3.51	3.55	3.58	3.59
Westerly	2.28	.....	.....	3.27	3.36	3.41	3.45	.....
Gneiss, Pelham 2 (∥)	1.84	.....	.....	3.24	3.46	3.58	3.63	.....
4 (⊥)	1.73	2.40	2.73	3.15	3.32	3.45	3.51	3.55
Syenite, Ontario 1	2.48	3.06	3.17	3.28	3.33	3.36	3.37	3.38
2	2.39	.....	.....	3.24	3.29	3.32	3.33	3.34
Norite, Sudbury 1	3.16	.....	.....	3.68	3.70	3.72	3.74	3.75
2	3.15	.....	.....	3.54	3.57	3.60	3.62	3.65
Diabase:								
Vinal Haven 3 (see text)	2.75	.....	.....	3.44	3.66	3.79	.....	.....
5	3.28	.....	.....	3.84	3.87	3.89	3.90	3.91
7	3.76	.....	.....	3.807	3.821	3.833	3.840	3.846
Maryland 1	3.60	.....	.....	3.75	3.77	3.79	3.81	3.83
3	3.73	.....	.....	3.78	3.80	3.81	3.82	3.83
Noranda 5	3.52	.....	.....	3.65	3.71	3.75	.....	.....
Metadiabase 3	3.97	.....	.....	3.98	4.00	4.01	4.02	4.03
Metadiabase 7	3.85	.....	.....	3.87	.....	.....	.....	3.89
Gabbro:								
Mellen 1	3.38	.....	.....	3.65	3.66	3.67	3.68	3.69
2	3.35	.....	.....	3.68	3.70	3.71	3.72	3.73
French Creek 1	3.39	.....	.....	3.84	3.90	3.94	3.96	3.98
2	3.15	.....	.....	3.82	3.91	3.94	3.96	3.98
Pyroxenite:								
Hypersthenite 1	4.52	.....	.....	4.56	4.57	4.58	4.59	4.60
2	4.45	.....	.....	4.48	4.49	4.51	4.52	.....
(New jacket) 2	4.47	.....	.....	4.51	4.52	4.54	4.55	4.56
Bronzitite	4.38	.....	.....	4.50	4.51	4.52	4.54	4.55
Dunite, Balsam Gap 1 (E)	4.13	.....	.....	4.42	4.44	4.48	4.51	4.53
2 (H)	3.81	.....	.....	4.47	4.51	4.56	4.59	4.61
3(H?)	4.37	.....	.....	4.57	4.60	4.63	4.64	4.65
(4 ⊥3) 4 (E?)	4.17	.....	.....	4.32	4.37	4.43	4.46	4.48
Obsidian, Modoc	3.52	.....	.....	3.51	.....	.....	.....	3.47

observed points lie on smooth curves with about the precision of the oscillator setting—about 1 or 2 parts in 3,000. At lower pressures the velocity is often uncertain by 1 or 2 per cent, due to irreversible effects, and especially at 1 atmosphere it is subject to very large fluctuations, depending upon the history of the sample. Thus, while any individual measurement at 1 atmosphere will be accurate to about 1 or 2 per cent, this velocity can easily be changed by as much as 20 per cent in some cases, so that the actual value given has little general significance. Above 1,000, not only is there very little hysteresis or irreversibility, but samples which have been maltreated in various ways tend to return to very nearly the same velocities at high pressure, regardless of their initial velocity. In a few cases closed loops were observed, the maximum width of loop being about 1 per cent of the condenser setting. Heating the specimens to 100° C. at high pressure usually produced a small permanent increase of velocity, generally of the order of several parts in 1,000. The effect of this might persist down to low pressures, or might not. Runs in which larger permanent changes than this occurred were not used for computing temperature coefficients.

In the tables the designation (H) signifies that the sample was cut with its axis in the plane of "hardest splitting"; similarly, (E) designates samples having the axis in the plane of "easiest splitting." The numbering of such specimens as had been previously used by Zisman was retained. Most of the samples were 19.7 cm long and 1.6 cm in diameter, but some were either shorter or greater in diameter. Thus the following were about 2.2 cm in diameter: diabases, Vinal Haven 7, Maryland 3, all of the Noranda specimens; hypersthenite 2, dunites 3 and 4. The following were notably shorter than 20 cm: bronzitite, 10.9 cm; dunite 1, 11.0 cm; dunite 3, 15.2 cm; dunite 4, 14.4 cm. The Balsam Gap dunite is very sugary and full of flaws; for the careful handling of this material we are indebted to MM. Wesley and Camille Fuller of the Quincy Granite Company, who prepared these and nearly all the other specimens. Of the Transvaal bronzitite, we had only a small hand specimen; otherwise it would have been possible to make specimens of nearly any size from this exceedingly strong, compact rock. Of the two samples of gneiss, 2 has its axis parallel and 4 has its axis perpendicular to the schistosity.

In several cases of Table 4 two sets of figures are given for a single specimen; these are distinct runs, sometimes with different jackets, and hence different jacket corrections. In the other tables a single run only is used for each of these rocks.

The Vinal Haven diabase 3 is a specimen that had been heated in air at 1 atmosphere to 800° C.; before this treatment the velocity in this specimen was about 3.7 km/sec; afterward it was exposed to high pressure, about 9,000 atmospheres, uncovered, and then to the same pressure, covered. This did not restore the velocity at 1 atmosphere to its former value, and in fact all the velocities are a little low, but only about 1 per cent at 2,000 atmospheres (with references to 4). The jacket leaked at the next pressure, and the experiment was terminated. This case illustrates the efficacy of pressure applied through a jacket in restoring even very badly shattered specimens to a state resembling the original one, at least with regard to velocity. The data for this specimen (3) were not included in the determination of average values.

In Table 5 are given the rigidities as function of pressure, in units of 10^{11} dynes/cm^2; these are simply calculated from the velocities and densities according to the expression $\mu = v^2\rho$. No correction is applied for the change of density; since the correction was not applied to the velocity, it should not be applied to the density here, as the two corrections nearly cancel out; the error is of the magnitude of the change of length under pressure, hence, again smaller than the other uncertainties.

Table 6 gives the average pressure coefficients of velocity and of rigidity for the three pressure intervals, 1,000–2,000, 2,000–3,000, and 3,000–4,000. The expressions for these coefficients have already been given; the change of length with pressure has been taken into account. The average pressure coefficient of velocity between two pressures is the difference of velocity at the two pressures divided by the velocity at the lower pressure and by the difference in pressure, here 1,000 kg/cm^2. It cannot be obtained with the precision of the measurements from Table 4, since too few significant figures are given. These coefficients are all positive.

The temperature coefficients of velocity of Table 7 are negative for all the rocks, that is, the velocity decreases in all cases when the

TABLE 5

RIGIDITY AS FUNCTION OF PRESSURE, AT 30° C

MATERIAL	RIGIDITY IN UNITS OF 10^{11} DYNES/CM^2			
	Pressure, Kg/Cm^2			
	1	500	1,000	4,000
Limestone, Solenhofen	1.96	2.20	2.30	2.47
Marble, Vermont 3	1.57	3.18	3.27	3.33
Quartzitic sandstone 2 (E)	3.06	3.98	4.23	4.42
3	3.14	3.82	3.97	.....
Granites:				
Rockport (H)	1.68	3.09	3.25	3.42
(E)	1.72	3.02	3.17	3.30
Quincy (100 ft.) 1	2.02	3.15	3.34	3.45
(235 ft.) 4 (H)	1.68	3.14	3.31	3.46
5 (E)	1.70	3.13	3.25	3.41
Westerly	1.37	2.83	2.99	.....
Gneiss, Pelham 2	0.90	2.77	3.16	.....
4	0.79	2.62	2.91	3.34
Syenite, Ontario 1	1.71	3.00	3.08	3.19
2	1.58	2.91	3.00	3.10
Norite, Sudbury 1	2.87	3.84	3.93	4.04
2	2.85	3.59	3.65	3.81
Diabase:				
Vinal Haven 5	3.19	4.37	4.42	4.53
7	4.17	4.30	4.32	4.39
Maryland 1	3.91	4.24	4.28	4.42
3	4.20	4.32	4.37	4.42
Noranda 5	3.70	3.99	4.11	.....
Metadiabase 3	4.78	4.81	4.84	4.93
7	4.44	.....	4.47	4.51
Gabbro:				
Mellen 1	3.32	3.86	3.89	3.95
2	3.25	3.93	3.97	4.05
French Creek 1	3.48	4.46	4.62	4.80
2	3.01	4.43	4.63	4.80
Pyroxenite:				
Hypersthenite 1	6.68	6.81	6.84	6.92
2	6.53	6.66	6.68	6.80
Bronzitite	6.28	6.65	6.66	6.80
Dunite, Balsam Gap 1 (E)	5.60	6.40	6.45	6.71
2 (H)	4.76	6.54	6.66	6.94
3	6.24	6.81	6.91	7.06
4	5.70	6.12	6.24	6.58
Obsidian, Modoc	3.03	.....	.....	2.97

TABLE 6

AVERAGE PRESSURE COEFFICIENTS OF TORSIONAL VELOCITY AND OF RIGIDITY, AS FUNCTION OF PRESSURE, AT 30° C

MATERIAL	PRESSURE INTERVAL, KG/CM²					
	$\left(\frac{1}{v}\frac{\Delta v}{\Delta p}\right)_i \cdot 10^6$			$\left(\frac{1}{\mu}\frac{\Delta \mu}{\Delta p}\right)_i \cdot 10^6$		
	1,000 to 2,000	2,000 to 3,000	3,000 to 4,000	1,000 to 2,000	2,000 to 3,000	3,000 to 4,000
Limestone, Solenhofen	23.0	7.4	3.9	48.0	17.0	10.0
Marble, Vermont 3	4.8	1.0	0.1	11.1	3.5	1.5
Quartzitic sandstone 2	7.5	4.8	2.9	17.7	12.3	8.5
3	11.1	4.0	.....	24.9	10.7	.....
Granite:						
Rockport (H)	13.6	6.9	1.5	29.3	15.6	4.8
(E)	14.9	1.7	1.7	31.9	5.2	5.2
Quincy (100 ft.) 1	15.6	6.7	2.3	33.3	15.2	6.4
(235 ft.) 4	10.2	6.5	5.3	22.2	14.8	11.4
5	10.2	6.5	4.1	22.2	14.8	10.0
Westerly	16.4	10.2	.....	34.6	22.2	.....
Gneiss, Pelham 2	33.6	15.9	.....	68.6	33.6	.....
4	38.4	16.9	9.7	78.6	35.6	21.2
Syenite, Ontario 1	9.3	1.8	1.8	20.0	5.4	5.4
2	10.1	2.6	0.8	21.7	6.7	3.1
Norite, Sudbury 1	6.1	4.3	2.4	13.4	9.8	6.0
2	6.9	6.8	6.8	15.0	14.8	14.8
Diabase:						
Vinal Haven 5	3.9	2.8	1.7	9.0	6.8	4.6
7	2.9	1.3	1.0	7.0	3.8	3.2
Maryland 1	5.1	5.1	3.9	11.4	11.4	9.0
3	2.9	1.8	1.2	7.0	4.8	3.6
Noranda 5	10.9	.....	.....	23.0	.....	.....
Metadiabase 3	3.8	2.1	2.1	8.8	5.4	5.4
7	1.6	1.3	0.8	4.4	3.8	2.8
Gabbro:						
Mellen 1	1.8	1.8	1.8	4.8	4.8	4.8
2	4.0	2.4	1.8	9.2	6.0	4.8
French Creek 1	8.1	4.8	3.7	17.1	10.8	8.6
2	9.7	3.8	2.5	20.6	8.8	6.2
Pyroxenite:						
Hypersthenite 1	1.9	1.7	1.5	4.8	4.1	3.9
2	2.7	1.7	1.6	6.3	4.3	4.1
Bronzitite	3.6	2.4	1.8	8.1	5.7	4.5
Dunite, Balsam Gap:						
1 (E)	8.2	6.5	3.9	17.3	13.9	8.7
2 (H)	10.9	4.5	3.6	22.7	9.9	8.1
3	4.4	2.8	1.6	9.7	6.5	4.1
4	12.9	6.6	3.8	26.7	14.1	8.5
Obsidian, Modoc	.....	− 3.95	.....	.....	−5.15	.....

temperature is increased from 30° to 100°. The pressure was maintained at 4,000 kg/cm² in all cases except one—Quincy granite 5—where it was 1,000 kg/cm². A correction for the change of length

TABLE 7

AVERAGE TEMPERATURE COEFFICIENT OF TORSIONAL VELOCITY FROM 30° TO 100° C., AT 4,000 KG/CM²

MATERIAL	$-\frac{1}{v}\frac{\Delta v}{\Delta t}\Big)_p \cdot 10^6$	PERMANENT INCREASE OF VELOCITY, AT 4,000 KG/CM² AND 30° C., DUE TO HEATING
		Per Cent
Marble, Vermont 3	127	0.13
Quartzitic sandstone 2	9	.10
Granite:		
Rockport (H)	37	.03
Quincy (233 ft.) 5	43 (at 1,000 kg/cm²)	.....
Norite, Sudbury 1	160	.06
Diabase:		
Vinal Haven 7	68	.....
Maryland 3	49	.09
Noranda 7 (metadiabase)	68	.....
Gabbro:		
Mellen 2	116	.02
French Creek 2	87	.31
Pyroxenite:		
Hypersthenite 2	76	.....
Bronzitite	102	.13
Dunite, Balsam Gap 1	208	.12
3	149	.08
4	127	.05
Pyrex glass	−57	.05
Obsidian, Modoc	−31	0.095 (decrease)

with temperature was applied; this is not known with precision, even at ordinary pressure; it is relatively small, however, except for the quartzitic sandstone whose velocity changes so little and, except for this case, introduces an uncertainty on the coefficient no greater than a few per cent. Pyrex and obsidian show an increase of velocity with

increasing temperature; this is also found at 1 atmosphere for these glasses and for pure silica glass.

Table 8 gives very rough values for the damping in a few of the samples expressed as the ratio of the resonant frequency (f_o) to the

TABLE 8

EFFECT OF PRESSURE ON DAMPING*

MATERIAL	$p = 200$ KG/CM^2	$p = 4{,}000$ KG/CM^2
	Q	Q
Limestone, Solenhofen	270	630
Marble, Vermont 3	140	1,000
Granite:		
Rockport (E)	180	1,700
Quincy (100 ft.) 1	180	1,200
Gneiss, Pelham 4	100	520
Syenite, Ontario 1	.....	2,000
Norite, Sudbury 2	240	330
Diabase, Vinal Haven 5	240	360
Gabbro:		
Mellen 1	340	1,700
French Creek	400	1,400
Dunite 2	400	500
Pyrex glass	2,200	.....

* Q is the cotangent of the loss angle, and is given by $f/\Delta f$, where f is the resonant frequency, and Δf is the width of the resonance curve at 0.7 the maximum ordinate; see text.

width of the resonance curve (Δf_o) at a certain amplitude. We should take the width of the curve at an amplitude equal to .707 times the maximum amplitude; the background of direct pickup from the driving unit and of stray induction from other sources and the departure from perfect symmetry of the resonance curve make any precise determination impossible. The viscous damping due to the nitrogen, and loss of energy to the supports, are also included; but for materials with so much internal damping, these effects are not

important. The curves of pure metals, for example, are very much narrower in spite of the same external losses. The value Q gives the ratio of elastic to dissipative force at the resonant frequency. R. L. Wegel and H. Walther[19] have shown that for metals and a few glasses Q is nearly independent of the frequency. The precision of our measurements of damping is not sufficient to permit any definite statement about the variation of damping with frequency, although very roughly, for rocks also, Q appears to be independent of the order of the harmonic.

TABLE 9

AVERAGE TORSIONAL VELOCITIES OF CERTAIN ROCK TYPES, AS FUNCTION OF PRESSURE, AT 30° C

MATERIAL	VELOCITY IN KM/SEC					
	Pressure, Kg/Cm²					
	1	500	1,000	2,000	3,000	4,000
Granite (7 samples)	2.55	3.42	3.50	3.54	3.57	3.58
Gneiss (2 samples)	1.78	3.20	3.39	3.51	3.57	3.61
Syenite (2 samples)	2.43	3.26	3.31	3.34	3.35	3.36
Diabase, metadiabase, and gabbro (11 samples)	3.53	3.78	3.82	3.84	3.86	3.87
Pyroxenite (3 samples)	4.46	4.51	4.53	4.55	4.56	4.57
Dunite (4 samples, all from Balsam Gap)	4.12	4.44	4.48	4.53	4.55	4.57

Finally, in Table 9 are gathered the average torsional velocities as function of pressure for several of the rock types represented here. There are clearly too few representatives of the syenite clan, for which we have only two samples of a single rock. The same is true of the gneiss, which, however, agrees very well with our average granite. As for the granite, it appears that the Rockport-Quincy type of granite is adequately sampled; the possibility of marked differences with other granites is indicated by the lone Westerly sample. This Westerly specimen does not seem, however, to be representative of the main body of Westerly granite, judging from its

[19] "Internal Dissipation in Solids for Small Cyclic Strains," *Physics*, Vol. VI (1935), p. 141.

low quartz content. It seems desirable to study additional granites of the other main types.

The spread of values among the diabases and gabbros which have been averaged together is greater than for the granites, and it is more difficult to say just what the average values represent. They approach very closely the values for the Maryland diabase, with the extreme deviations approximately ± 4 per cent from the mean. This is rather less than the spread of compressibility even at 10,000 atmospheres observed for several diabases by Adams and Gibson,[20] who concluded that the effects of structure may persist in basic rock even to 10,000 atmospheres. There are some indications that somewhat greater uniformity would be produced by a pressure of perhaps 10,000 kg/cm^2; but an examination of Table 6 in conjunction with Table 4 will show that a good deal of scattering must persist, since some of the rocks with the highest velocities have also high pressure coefficients of velocity. Within this group there is a fairly good correlation of velocity with composition, the rocks with the smaller proportions of feldspar and higher proportions of pyroxene and olivine having the higher velocities, except that the positions of the Vinal Haven and Maryland diabases are apparently reversed. When we attempt to account for such small differences, however, we require a more directly pertinent analysis than is at present available; thin sections are not generally representative enough to furnish precise averages for large specimens; this is, of course, especially true when, as in the present case, the thin sections were not cut from the specimens themselves but from other portions of the samples from which the specimens were made. Finally, there is the possible influence of crystal size. The longest, if not the bulkiest, crystals are those of the Mellen gabbro and the syenite. In both cases, however, the differences, between different samples is only about 1 per cent. Such agreement would be unlikely if individual crystals played an important part in determining the velocity. It is not impossible that one kind of crystal—in these cases the feldspars, which are by far the longest—plays a disproportionate part in determining the velocity. For general purposes the average values will probably be most pertinent,

[20] *Op. cit.*

with the proviso that individual specimens similar to those studied may deviate by at least as much as 4 per cent at high pressure, and, of course, very much more at low pressure.

The sampling of pyroxenite rocks is very limited, although the samples which we have studied are probably good representatives of nearly pure aggregates of hypersthene and of bronzite crystals. The difference, as one would expect, is small. In the case of the dunite, we have a single large block of the Balsam Gap material from which were made four samples, one thick one and one thin one, with their axes in the plane of "hardest splitting," and, similarly, one thick one and one thin one with their axes in the plane of "easiest splitting." The velocities appear to be quite definitely correlated with these directions, differing by about 2 per cent. The mean of these four is probably representative of this rock with at least this precision. Some differences due to structural variations will doubtless exist among rocks of the same chemical and mineralogical composition even at 4,000 atmospheres, and for this reason samples of other dunites should be studied if they can be obtained. Such alterations of structure as we have been able to produce by heating under pressure are extremely small for the dunite, but higher temperatures or higher pressures may prove more effective. The relatively high values of the pressure coefficient of velocity in the range 3,000–4,000 atmospheres for three of the four dunite samples suggests the persistence of some porosity at these pressures, and it is also noteworthy that the sample with the highest velocity (3) has the smallest pressure coefficient. Thus the velocity in perfectly compact dunite of this composition may be slightly higher than the average for these specimens, but it seems unlikely that the difference will exceed a few per cent.

FAULT BORDER OF THE SANGRE DE CRISTO MOUNTAINS NORTH OF SANTA FE NEW MEXICO

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ABSTRACT

The eastern prong of the southern Rocky Mountain province in New Mexico, called the Sangre de Cristo Mountains, is usually described as a great anticlinal structure. Its eastern border and general internal structure conform to this definition but its western border is a fault-line scarp facing the down-faulted structural basin of the Rio Grande depression.

This depression is underlain by partly consolidated fan deposits—the Santa Fe formation of Miocene or early Pliocene age. A water-laid tuff, the Picuris formation, with interbedded basalt in places underlies the Santa Fe unconformably. These two formations are separated from the older rocks of the Sangre de Cristo both by erosional and by fault contacts. The fault pattern is irregular and consists largely of en echelon faults trending northwest, but so overlapping as to produce a north-south trend of the scarp from Santa Fe to Chimayo. North of this point the Picuris re-entrant forms an extension of the depression eastward, and the Picuris prong, bounded by a complex fault pattern, forms an extension of the mountain area westward.

The Rio Grande depression has been eroded in several successive stages. At the earliest stage the area of outcrop of the Santa Fe was reduced to a plain and the fault scarp was reduced to maturity by erosion. Later revival of streams has accentuated the topography and given the mountain front an apparent height about twice as great as it had in the early stages of erosion.

LOCATION OF THE AREA STUDIED

The area here described lies on the west flank of the Sangre de Cristo Range in the north-central part of New Mexico (Fig. 1) between the parallels 35°40′ and 36°20′ and meridians 105°35′ and 106°00′. It occupies the northern part of Santa Fe County, southeastern Rio Arriba County, and southern Taos County. In terms of the land-office surveys it is bounded by R. 9 to 12 E. of the New Mexico principal meridian and T. 17 to 24 N. of the state base line. Physiographically the area lies partly in the Rio Grande depression and partly on the eastern prong of the southern Rockies, both of which are within the southern Rocky Mountain province as defined by Fenneman.[1]

Most of the Rio Grande depression north of Santa Fe has "bad

[1] N. M. Fenneman, "Physiographic Divisions of the U.S." *Annals Assoc. Amer. Geog.*, Vol. XVIII (1928), pp. 328–31.

land" topography, but there are also broad sloping mesas and ridges that rise eastward to the mountain mass. These are remnants of erosion surfaces that once spread as smooth plains outward from the mountain front. The border of the mountains rises from these dissected plains in relatively gentle spurs and ridges, presenting a subdued and mature front.

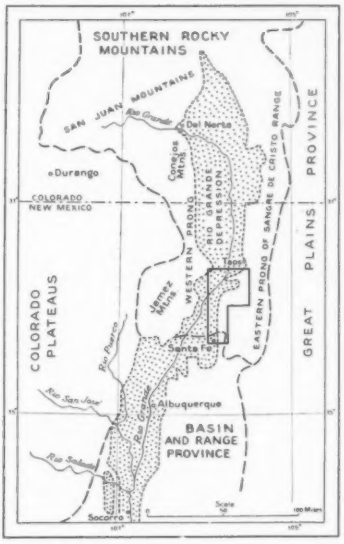

FIG. 1.—Index map showing relation of the area studied to the Rio Grande depression and the physiographic provinces in north-central New Mexico and southern Colorado.

The Sangre de Cristo Mountains begin in Colorado on the south flank of the Arkansas Valley near Salida, and extend southward for 250 miles. As shown diagrammatically in Figure 1, the western border of the range is marked by westward projections and eastward re-entrants. South of the town of Taos there is a sharp projection westward, the Picuris prong, in which pre-Cambrian gneiss and schist extend to the Rio Grande. South of the prong lies the Picuris re-entrant, which is 20 miles from east to west and 5 miles from north to south. On the opposite side of the Rio Grande and separated from the westward highland by faults is the Abiquiu re-entrant, which has been described by H. T. U. Smith.[2] South of the Picuris re-entrant the western front of the mountains has a nearly north-south trend, with only minor irregularities to a point south of Santa Fe. The portion of this long western border from Santa Fe north, including the Picuris prong, is the area considered in this paper. The range east of this area is divided into two highland masses separated by the north-south Pecos Valley. The elevation of the peaks in the western highland, east of the city

[2] "The Tertiary and Quaternary Geology of the Abiquiu Quadrangle, New Mexico" (Doctor's thesis, Harvard University, 1935). Pp. 231.

of Santa Fe, is about 12,000 feet, but the crest becomes lower to the south, and at the Glorieta Pass the range ends at elevations of about 7,000 feet.

ROCKS OF THE SANGRE DE CRISTO MOUNTAINS

Larger areas of pre-Cambrian crystalline rocks occur in the Sangre de Cristo Range, although the areas shown on the United States Geological Survey's geologic map of New Mexico are somewhat exaggerated. It is known that, in general, granite gneiss with minor masses of basic rocks is intruded into schist. Two areas which stand out as different from most of the pre-Cambrian mass are the Picuris prong and the area between the En Medio and Chupadero creeks, 2 miles east of the Santa Fe National Forest fence. At the first locality the schist is characterized by staurolite and sillimanite. Pegmatites with lepidolite and other relatively rare minerals also occur here at the Harding Mine.[3] In the second locality massive pegmatite crops out over a large area.

The Pennsylvanian Magdalena limestone, from 900 to 2,500 feet thick,[4] lies unconformably on the pre-Cambrian and is the surface rock over large areas in the mountains. Limestone is the predominating rock, but sandstone and shale also occur interbedded.

Permian, Jurassic, Cretaceous, and Eocene rocks occur in localities south and west of Santa Fe but are absent on the western slope of the mountains either because they were not deposited or because they have been later removed.

The assumption heretofore held that the mountains owe their form and elevation to folding is based on the attitude of the Magdalena, which is strongly folded and faulted and lies in great belts parallel to similar belts of pre-Cambrian rock having the characteristic north-south trend of the mountains. Folds involving Cretaceous and doubtfully Eocene rocks also occur on the eastern border of the range in structures similar to those of the Front Range in Colorado. These structural phenomena are, however, in marked contrast with the fault border on the west, which is the subject of this paper.

[3] W. T. Schaller and E. P. Henderson, "Purple Muscovite from New Mexico," *Amer. Mineral.*, Vol. XI (1926), pp. 5–16.

[4] N. H. Darton, "Red Beds and Associated Formations in New Mexico," *U.S. Geol. Surv. Bull. 794* (1928), p. 19.

PICURIS TUFF

The Tertiary basin deposits of the area include considerable thicknesses of tuff and conglomerate which are unlike the Santa Fe formation both in texture and in composition. These beds are tentatively regarded as of pre–Santa Fe age because of their known unconformable relation to the Magdalena formation and their close resemblance to the pre–Santa Fe beds of immediately adjacent areas. The name "Picuris tuff" is here given to these beds, which crop out typically in the Picuris re-entrant (Fig. 2) near the towns of Badito and Placita.

In the Badito locality, about 1,000 feet crop out in one continuous section. At the base there are 680 feet of buff water-laid tuff interbedded with cemented gravel containing a large percentage of pre-Cambrian material. Above lie 190 feet of salmon-pink, sandy clay with layers of cemented gravel containing many volcanic pebbles. These beds are followed by 10 feet of giant conglomerate characterized by rounded boulders of pre-Cambrian quartzite and of quartzose sandstone from the Magdalena formation, having a maximum size of about 4 feet. Above this bed there are 100 feet of gravel similar but with smaller pebbles and cobbles capped by 40 feet of basaltic lava. Some gravel occurs above the basalt but its thickness is not known. About 2 miles east of this section, red and green silts and clays and a giant conglomerate, totaling 200 feet in thickness, rest unconformably on the Magdalena. These beds lie below the tuff so that the total thickness of the Picuris in this area seems to be in excess of 1,200 feet.

At other localities in the area similar volcanic beds occur. Their relation to the Santa Fe is also obscure. Between Penasco and Chamizal (Fig. 2) 50 feet of breccia and conglomerate are exposed. Part of the outcrop is covered by landslides. Here the presence of volcanic pebbles in a disturbed zone seems to indicate that the tuffaceous material may be separated from the Santa Fe beds by a fault. Tuff beds, with interbedded basalt flows, crop out at irregular intervals along the mountain front between En Medio and the city of Santa Fe. They are closely associated with the fault-line scarp and in places intervene between the normal Santa Fe beds and the older rocks in such a position as to imply unconformity of the Santa Fe on the Picuris.

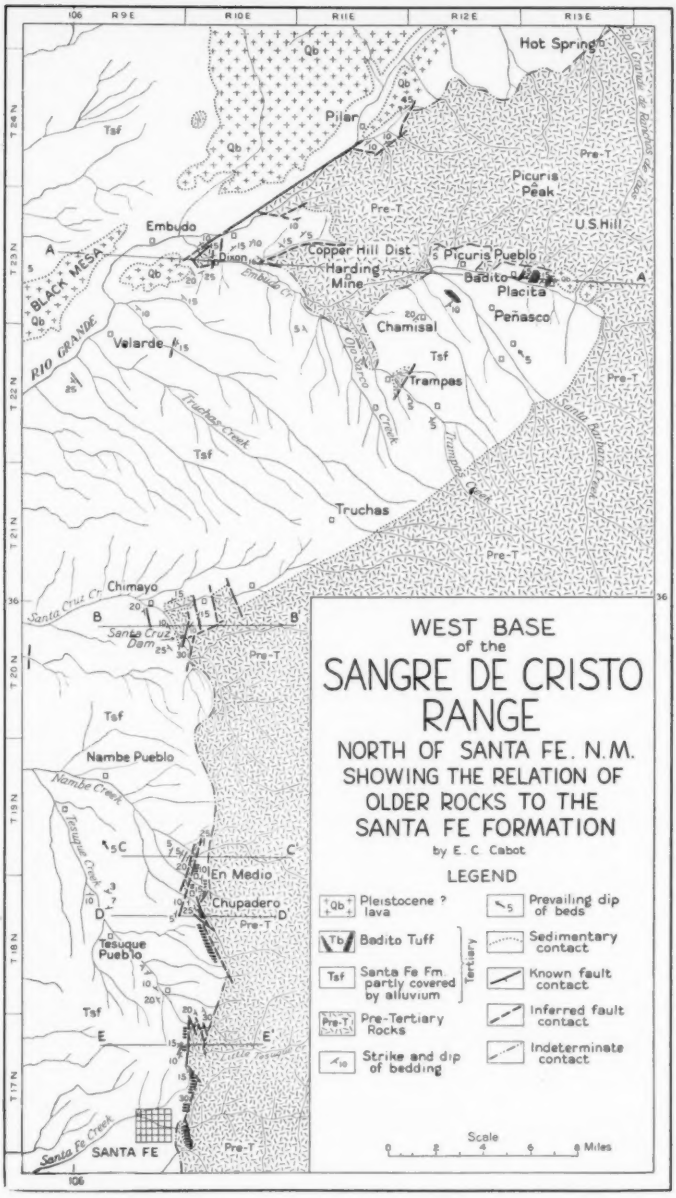

FIG. 2.—Geologic map of an area on the west flank of the Sangre de Cristo Range north of Santa Fe, N.M.

SANTA FE FORMATION

The Rio Grande depression is underlain by partly deformed alluvial beds, called the "Santa Fe formation." In this area the Santa Fe is a semiconsolidated fan deposit which is coarser on the east and north adjacent to the mountains and gradually becomes finer to the west and south. Many of the beds are salmon pink in color and the general aspect is reddish. Buff, white, and gray beds exist. The beds, as a whole, represent fan deposits which, however, are much coarser and present greater variety in texture and in composition than the fine-grained fan deposits of the Ceja del Rio Puerco[5] which were deposited by ephemeral and intermittent streams originating in an area of sedimentary rocks. In contrast, the beds of this area were laid down by similar streams originating in the pre-Cambrian areas of the Sangre de Cristo Range. The disturbed conditions of the beds and the lack of horizon-markers make determinations of thickness difficult, but it is obvious that the previous estimates of 1,200–1,500 feet are conservative (Figs. 3 and 4). From recent extensive collections of vertebrate fossils, Childs Frick,[6] whose studies are still in progress, believes that the beds range in age from Middle Miocene to Lower Pliocene.

PEDIMENTS

The deformed Santa Fe beds have been subjected to erosion at several different times, the earliest perhaps dating back to late Pliocene time. At each stage the streams flowing out of the Sangre de Cristo Mountains cut broad pediments graded to the Rio Grande. Although developed largely on the Santa Fe and Picuris formations, the pediments are also eroded on the rocks of the mountains and obliterate the original differences in elevation due to faulting. Remnants of these pediments occur at various levels (Figs. 5 and 6).

The highest and presumably the oldest of these pediments is best preserved as a broad, gently sloping surface cut on the crystalline rocks of the mountains. Where this surface cuts the fault line at the

[5] Kirk Bryan and F. T. McCann, "The Ceja del Rio Puerco: A Border Feature of the Basin and Range Province in New Mexico," *Jour. Geol.*, Vol. XLV (1937), pp. 801–28.

[6] "New Remains of Trilophodont—Tetrobelodont Mastodons," *Amer. Mus. Nat. Hist. Bull. 59* (1933), pp. 549; "The Horned Ruminants," *Amer. Mus. Nat. Hist. Bull. 69* (1937), pp. 669.

FIG. 3.—Deformed Santa Fe formation on Embudo Creek: pre-Cambrian floor exposed in foreground; in the background high pediment east of Dixon on the left and Velarde Mesa on the right.

FIG. 4.—Deformed Santa Fe formation in Dixon sub-basin

base of the mountains it has an elevation of approximately 8,000 feet and may be 1,000 feet above the present stream grades. The surface is also preserved on a few gravel-capped remnants in the vicinity of the town of Truchas, extending westward for about 11 miles, with an average grade of 87 feet per mile. This grade reaches

FIG. 5.—Mountain front east of Santa Cruz dam showing beveled surface on pre-Tertiary rock (No. 1) and gravel-capped pediment (No. 2) that is possibly equivalent to the Ortiz pediment.

the Rio Grande near Dixon at an elevation about 1,100 feet above the present grade of the river. (The lava-capped mesa between Dixon and Velarde has about the same elevation as the westernmost remnant of this pediment.) The pebbles in the gravel of the highest pediment are commonly 2 inches to 1 foot in diameter. However, there are also larger, well-rounded boulders up to 6 feet in diameter, many of which have a surface blackened by desert varnish. These gravels extend across the lines of faulting between the Santa Fe beds and the rocks of the mountains and present a formidable obstacle to structural study.

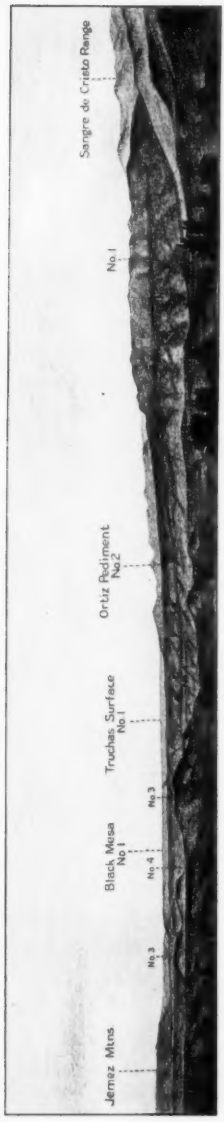

FIG. 6.—Fault-line scarp of the Sangre de Cristo Range: view north from southern border of Tesuque Valley sketched from photographs (No. 1, early pediment; No. 2, Ortiz pediment; Nos. 3 and 4, later partial pediments).

The age of the highest pediment cannot be conclusively determined. If the basalts near Pilar and Velarde are of Pliocene age, then this surface is also of Pliocene age. The locality near the town of Truchas is so close to the mountain that it seems possible that, if the mountain glaciers during the Pleistocene were extensive, they might have covered it. No till or other undoubted evidence of the presence of ice was found, but there are bodies of iron-stained sand and clay that might be glacial outwash. It is possible that by tracing this highest surface or perhaps lower ones into the mountains, some connection might be found between them and the glacial history of the region. Certainly such an inquiry seems to be the best hope of attaining a date for the sequence of successively lower pediments.

Several streams in this area are clearly superimposed. Embudo Creek and several of its tributaries (Fig. 2) head in the resistant crystalline rocks, flow onto the relatively weak beds of the Santa Fe formation and the Picuris tuff, then flow again across the crystallines in the vicinity of the Harding Mine. The north and south branches of Santa Cruz Creek (Fig. 2) behave similarly. These streams must have been let down from a cover of Santa Fe beds and probably from erosion surfaces of post–Santa Fe date. This relationship, the scarcity of remnants of

the highest pediment, and the height of the fault-line scarp of the Sangre de Cristo Mountains point to the removal of a tremendous amount of detritus from this part of the Rio Grande depression.

Between Santa Fe Creek and Little Tesuque Creek, north of the city of Santa Fe, there is a high, gravel-covered pediment remnant that forms the ridge followed by the Circle Drive. In this vicinity it has an altitude of 7,500 feet and slopes westward toward the Rio Grande, where it has an altitude about 500 feet above the river. Patches of gravel along the mountain front as far north as Chimayo have about the same altitude. These remnants of gravel seem to represent a continuation of this pediment. Near the mountains it is from 500 to 600 feet lower than the Truchas pediment and appears to represent a lower grade of the river. It is indicated as the No. 2 surface in Figure 6 and may be the equivalent of the Ortiz pediment identified by Bryan and Upson[7] in the Santo Domingo area.

In the Tesuque Valley there are two lower erosion surfaces indicated as No. 3 and No. 4 (Fig. 6) below the supposed Ortiz pediment, and there is also a still lower terrace not shown in this figure. In other valleys, notably the Embudo Valley, there appear to be as many as four or five terraces. The correlation of these terraces with one another is impossible without further study, and attention is called to them in this connection chiefly because the gravel of the terraces is a major factor in concealing the underlying Santa Fe beds.

GENERAL DESCRIPTION OF THE WESTERN BOUNDARY OF THE MOUNTAINS

The structural problem of the area is a part of the larger problem of the relation of the Rio Grande depression to the adjacent highlands. Here, as elsewhere, the basin has been carried down relative to the highlands. However, only a small part of the depression is shown on Figure 2, and much of the structural detail in this limited area is concealed beneath a gravel cover. This discussion is thus confined to the eastern border of the basin and does not consider the faults wholly within the basin deposits. At only a few places can these faults be recognized, and none of them have been followed for

[7] Kirk Bryan and J. E. Upson, "Geomorphology of the Santo Domingo Area, New Mexico" (unpublished MS).

any distance because of the lack of horizon-markers. A large part of the general movement may have taken place on these unmapped faults.

The nature of the contact between the Tertiary and the older rocks at the western base of the Sangre de Cristo Mountains is the chief structural problem considered. Between the city of Santa Fe and the village of Chimayo, for a distance of 20 miles, the contact between the Santa Fe beds and the older rocks is a nearly north-south fault contact (Fig. 2). It is marked, however, by recesses formed by en echelon faulting and by outlying fault blocks of pre-Tertiary rock. Its general character is in marked contrast to that of the next section to the north, from Chimayo via the village of Truchas to a point about 6 miles east of the pueblo of Picuris. At the sharp bend near Chimayo there is evidence of strong faulting, but where the trend is northeasterly the contact may be depositional. The Santa Fe beds are not well exposed, but wherever observed they have dips of about 5° to the northwest and thus away from this contact. Apparently the contact is depositional and has been so shown on Figure 2.

The Picuris re-entrant is bounded by the northeasterly contact just described and an irregular, generally east-west, contact which forms the south boundary of the Picuris prong. On general grounds this contact should be a fault, as the Santa Fe beds dip toward the contact, but local evidence of faulting is difficult to obtain. From a point near Badito to the head of the canyon of Embudo Creek, west of Picuris Pueblo, the variation in strike and the considerable dip of the Picuris tuff and Santa Fe beds compel the assumption of faulting. Along the rest of this portion of the contact the Santa Fe beds may rest unconformably on the older rocks, but, if so, their present dip would indicate that they were laid against the slopes of a rather steep hill.

Within the Picuris prong, east of the town of Dixon, is a deformed graben, bounded by faults north and south, with a depositional contact on the east. West of the town is an isolated block of pre-Cambrian rock completely surrounded by faults.

Northwest of the Picuris prong there is evidence of strong faulting.

Here the Rio Grande flows parallel to the major fault of the area and, except at the southwestern end, there are no outcrops of the Santa Fe formation, for a maze of landslides, involving both the lava from the mesa to the west and the soft Santa Fe beds, covers the western slope of the canyon of the Rio Grande. On Pilar Creek, southeast of the town of Pilar, there is a small recess bounded by faults on the south and east. From the northerly end of this recess the northern border of the Picuris prong rises as hills above the smooth, depositional plains of the Taos re-entrant. Here the contact is concealed and its nature speculative.

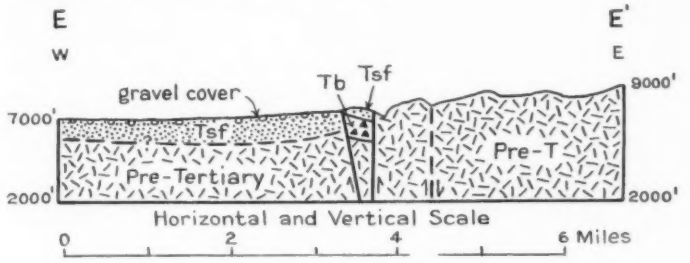

FIG. 7.—Geologic section *E–E'* on Little Tesuque Creek

DETAILS OF THE FAULT SYSTEM

The character of the faulting at the contact of the Santa Fe beds and the pre-Tertiary rocks of the mountain block can best be understood by a consideration of the details of faulting at selected localities.

On Little Tesuque Creek, as shown in section *E–E'* (Fig. 7), a well-defined fault with a northwest strike separates the Santa Fe formation from the Picuris tuff. West of the fault the Santa Fe is represented in the section as having a low westerly dip, but this is largely a conventional device. A consideration of the dips recorded to the north and west in the valley of Tesuque Creek (Fig. 2) shows that the beds are much disturbed and the structural pattern, whatever it may be, is very confused. There are probably faults of large throw in addition to numerous small faults of indeterminate throw, many of them filled with gypsum. East of the fault shown in the

section *E–E′*, the Santa Fe beds overlie the Picuris tuff and rest against the pre–Santa Fe rocks along a well-defined north-south fault. The pre-Tertiary rocks here are represented by the Magdalena limestone resting unconformably on the pre-Cambrian crystallines. To the east of this fault the limestone is offset by a normal fault which is the southern extension of the fault which borders the recess to the north on its northeast side.

Near the villages of En Medio and Chupadero the faulting is complex, and a narrow block of pre-Tertiary rocks has been uplifted along a strong northeasterly fault. The section *D–D′* (Fig. 8) shows the southern extension of this fault where its throw is moderate.

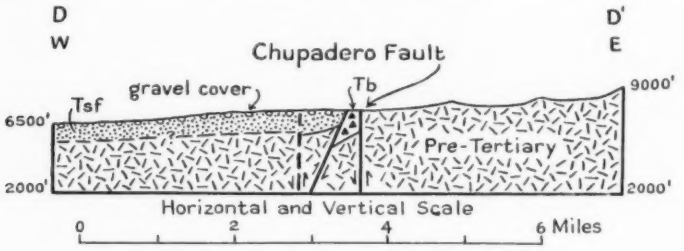

FIG. 8.—Geologic section *D–D′* near village of Chupadero

The section extends eastward to the mountain front, crossing first a recess between a sharply northwest fault and the Chupadero fault. In this recess the Picuris tuff, with interbedded basalt flows, has a dip of 25° SW., where it rests against the pre-Cambrian along a visible fault contact. The mashed beds at the contact form a soft fault breccia 5 feet thick.

The nature of the structure north of En Medio is shown in section *C–C′* (Fig. 9). The western end of this section begins with Santa Fe beds that dip gently westward. The dip changes from 5° W. to 5° E. on approaching a fault of small throw that lies wholly within the Santa Fe formation. Between this fault and the pre–Santa Fe rocks of the outlying block already mentioned the dip is 20° NW. Eastward the Santa Fe beds rest unconformably on the older rocks, dipping westward. Were the change in dip the sole criterion for faulting, it would be impossible to tell whether this were a fault or a

sedimentary contact. However, the strike of the beds is more northeasterly than the trend of the contact, and this discrepancy indicates a fault which may have a small throw. East of the block of older rocks the Santa Fe beds form a graben and dip 25° W., maintaining

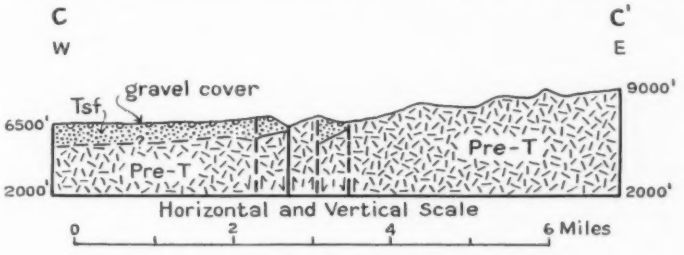

FIG. 9.—Geologic section *C–C'* near village of En Medio

this dip as far as the main mass of the mountain. This last-named contact is also a fault, and farther south the graben is floored with beds of Picuris tuff.

East of Chimayo and west of the great bend in the mountain front (Fig. 2) is a small outlying block of pre-Tertiary rock (granite

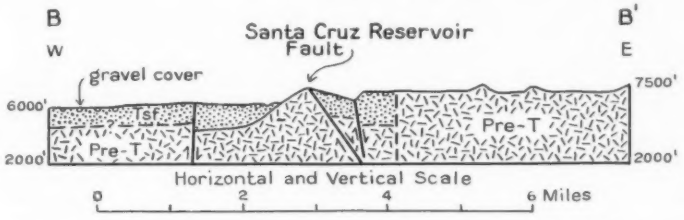

FIG. 10.—Geologic section *B–B'* near village of Chimayo and just north of Santa Cruz dam.

gneiss). Beginning on the west, as shown in section *B–B'* (Fig. 10), the Santa Fe beds dip westward and are broken by a small fault containing a sand dike. Eastward they rest unconformably on the block of pre-Tertiary material which forms the footwall of the Santa Cruz reservoir fault. This well-marked fault extends along the shore of the Santa Cruz reservoir north of the dam and gives rise to a bold, inclined wall of gneissic granite carrying patches and sheets of

granite breccia from 2 to 3 feet thick. It dips 55° E. Near the north end of the north arm of the reservoir the Santa Fe beds rest on the granite in fault contact, as shown in Figure 11. The Santa Fe here consists of pinkish, sandy beds dipping eastward away from the contact *F–F′* in Figure 11, but at a distance of 600 feet they dip 10° W. Along the fault there is a zone of mashed sand about 2 feet

FIG. 11.—Santa Cruz fault: face of granite fault breccia on left (*west*) and Santa Fe formation on right (*east*); contact at *F–F′*.

thick. Evidently the granite breccia was formed at a depth where granite rubbed on granite. It was then cemented with calcium carbonate and was not disturbed by the movement past it of the Santa Fe beds, which were mashed to form a second fault breccia. North of this locality there is another fault in en echelon relation to the Santa Cruz reservoir fault. Probably other complications lie concealed beneath the heavy gravel of the dissected pediment. In section *B–B′* this northerly en echelon fault is shown as extended southward, thus intervening between the Santa Cruz fault and the fault which bounds the main mountain mass.

The structure of the Picuris re-entrant is so concealed by the alluvium of the early and late epochs of erosion that a structure section here is likely to be inaccurate. In section *A–A′* (Fig. 12) a section from the Rio Grande Valley through the Dixon graben and the northern part of the Picuris re-entrant is presented. The attitude of the Santa Fe beds below the cover of basalt and landslide material at the western end of the section is conjectural, as exposures are rare except in the vicinity of the small fault block west of Dixon. The fault on the southwest side of the Dixon sub-basin is inferred from the wide angle between the strike of the beds and the trend of the contact. Crossing the southern part of the prong, the section passes through a re-entrant containing the Pueblo of Picuris. Here there are only a few outcrops of Picuris and of presumed Santa Fe beds, but it is believed that the dips are low and that the boundary on the north is a fault. East of this point the section passes through supposed Santa Fe formation to a fault separating it from the Picuris tuff, which, in turn, is partly faulted against and in part unconformable on the Magdalena limestone and the pre-Cambrian. The basalt which is shown in Figure 12 as overlying these beds unconformably is actually conformable and a part of the Picuris tuff.

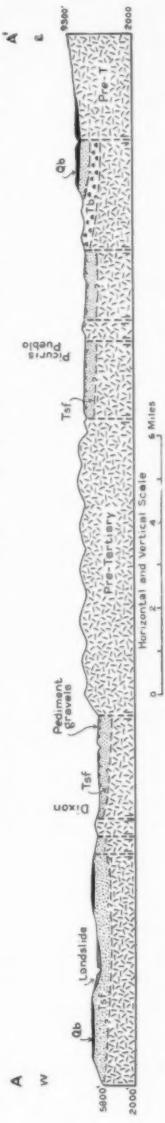

FIG. 12.—Geologic section *A–A′* from the Black Mesa eastward to the mountains through southern portion of Picuris Prong

SUMMARY ON THE FAULT-LINE SCARP

The Sangre de Cristo Mountains, viewed from the central portion of the Rio Grande depression, present a bold and rugged appearance. If seen from one of the higher pediments, they are relatively subdued in form and there is no scarp at the base of the mountains. In the

early stages of erosion the original fault scarp was obliterated and streams flowed indifferently across the hard rocks of the mountains and the soft rocks of the structural basin west of the main line of faulting. The relative elevation of the mountain block was, however, retained, and peaks rose from 3,500 to 4,000 feet above the pediments at the base. However, the dissection of the pediments by the present stream system has rejuvenated the fault-line scarp by removal of the less resistant Santa Fe, so that, in places, the mountains present a bold front, from 1,000 feet to 1,200 feet higher than the subdued scarp visible from the highest pediment remnants.

The deformation of the Santa Fe beds and their steep dips, whether away from or toward the contact with the older rocks, constitute general evidence of faulting. The occurrence of the Picuris tuff in small, tilted blocks between the general area underlain by Santa Fe beds and the great areas of older rocks that form the highland to the east indicates that it has been dragged up along the major fault or a series of faults. The confused pattern of the dips of the Santa Fe beds is proof of the complexity of the deformation. In a few places such as the Chupadero and Santa Cruz faults, the fault contact has been observed and studied. These localities are critical and here the proof of faulting is complete.

So far no evidence has been obtained as to the amount of displacement of the Rio Grande depression relative to the Sangre de Cristo Mountains. One might expect to estimate this from the difference in elevation of the base of the Santa Fe beds in the basin and that of the pre–Santa Fe surface on the mountain block, but unfortunately no direct observations can be made on the position of the base of the Santa Fe, nor do wells penetrate deeply enough to reach it. The mountain area has been insufficiently studied but no pre–Santa Fe surface has yet been found. Thus the amount of displacement is unknown, but it must be relatively large as the mountains rise from 3,000 to 4,000 feet above the highest outcrops of the Santa Fe.

The folded structure of the interior of the Sangre de Cristo Range developed during the Laramide orogeny gives no clue to the faulted character of its western border, which dates from a late Tertiary or early Pleistocene uplift.

ACKNOWLEDGMENTS.—The field work was done from 1932 to 1935 under the direction of Professor Kirk Bryan. I am much indebted to Mr. Charles S. Denny for help in the field during 1932 and to Dr. H. T. U. Smith for help during 1935. Mr. Franklin T. McCann has helped with the text and Mr. Edward Schmitz has prepared the drawings.

THE SALINITY OF THE OCEAN IN RELATION TO WATER VAPOR IN THE ATMOSPHERE AND THE LEVEL OF THE SEA

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ABSTRACT

Investigation shows that, if all the water vapor in the atmosphere were returned to the sea, sea-level would rise less than 1 foot. The increased salinity of the ocean from pre-Cambrian times to the present has had a negligible effect on the amount of water vapor in the atmosphere.

For a long time the blanketing effect of atmospheric water vapor on heat at the surface of the earth has been recognized as a factor in climatic changes. To play a part in any major variation of climate the amount of water vapor in the atmosphere would have to change and remain changed for a long period of time. The difficulty is in finding any cause for such variation in atmospheric water-vapor content, unless the relative extent of land and sea areas has changed sufficiently. Clark[1] has attempted to solve this problem by suggesting that an increase in the salinity of the ocean throughout geologic time would cause a decrease in the vapor pressure of the ocean and hence a decrease in evaporation from its surface. The net result would be, according to Clark, less water vapor in the atmosphere and a less uniform climate. Clark furthermore stated that the reduced evaporation would mean that more water remained in the ocean and that, therefore, the water level would become higher. He thought that this would throw some light on faunal migrations of the past, for

> if the present oceans were of fresh water their surface would be far below the present level, and numerous land connections would appear by which a general interchange of faunas now unconnected would be possible.

It is the purpose of this paper to test these hypotheses. First, let us consider how much change in ocean level would result if the

[1] Austin H. Clark, "A Comparative Study of the Most Ancient and the Recent Marine Faunas," *Jour. Wash. Acad. Sci.*, Vol. XIV (1924), pp. 487–91.

earth's atmosphere were to vary from one completely saturated with water vapor to one devoid of all moisture. This is the greatest variation which is possible.

According to Clarke,[2] the mass of the atmosphere is equivalent to 1,268,000 cubic miles of water. At 80° F., air saturated with water weighs .07253 pounds per cubic foot while the water vapor alone weighs .001576 pounds per cubic foot of saturated air.[3] Hence the maximum amount of water possible in the atmosphere at 80° F. is .0217 of the mass of the atmosphere, or 27,516 cubic miles of water. The surface area of the oceans is 139,440,000 square miles.[4] Therefore, if all the water in saturated air were returned to the sea, the total rise of sea-level would be only 1.04 feet. In any actual case the rise of sea-level due to decreased vapor pressure and evaporation must be much less than this.

For the second point, the actual change in vapor pressure between pure water and water having the salinity of the present oceans was calculated. Rough averages were computed for the salinity and constituents of ocean water from figures given by Clarke in "The Data of Geochemistry."[5] The salinity of ocean water is approximately 3.6 per cent, and the percentages of the constituents are as shown in Table 1. In terms of ion concentration, the total is approximately 1.149 per 1,000 grams of water. Applying Raoult's Law for the vapor pressure of a solution:

$$P = \frac{55.56}{56.71} p_o ,$$

where P is the vapor pressure of the solution and p_o is the vapor pressure of pure water, we get $P = .98p_o$.

Thus the total decrease in vapor pressure from an ocean totally lacking in dissolved salts to one having the present-day concentration is 2 per cent. In terms of millimeters of mercury, at 20° C. the saltless ocean would have a vapor pressure of 17.535 mm.[6] while the

[2] F. W. Clarke, "The Data of Geochemistry," *U.S. Geol. Surv. Bull. 770* (1924), p. 22.

[3] *Handbook of Chemistry and Physics* (19th ed., 1934), p. 905.

[4] *World Almanac* (1936), p. 786.

[5] Clarke, *op. cit.*, pp. 126 and 127.

[6] *Handbook of Chemistry and Physics* (19th ed.), p. 1228.

saline ocean of today would have a pressure of 17.184 mm. This is a change in pressure of .351 mm. of mercury. When one considers that a change in the atmosphere of from 60 to 80 per cent humidity is responsible for an atmospheric vapor-pressure change of about 3.5 mm. of mercury, the decrease due to salinity becomes insignificant. Moreover, if a 100 per cent vapor-pressure drop will cause sea-level to rise about 1 foot, then a 2 per cent vapor-pressure drop will modify sea-level only to the extent of about $\frac{1}{4}$ inch.

TABLE 1

CONSTITUENTS OF THE OCEAN WATERS
(IN IONIC FORM)

	Per Cent
Chloride	55.40
Bromide	0.18
Sulphate	7.75
Carbonate	0.20
Sodium	30.60
Potassium	1.20
Calcium	1.20
Magnesium	3.70
	100.23

The conclusion reached, therefore, is that the salinity of the ocean cannot play any important part in causing changes in either atmospheric water-vapor content or sea-level. This conclusion differs from the view expressed by Visher[7] that increased salinity has cut down the amount of water vapor in the air because the rate of evaporation was lessened. The question is whether rate of evaporation would be of any significance over a long period of time in which conditions between the atmosphere and the ocean would be able to reach equilibrium. It would seem that the vapor pressure of the ocean should be the important factor in limiting the amount of atmospheric water vapor. If this is true, then the effect of salinity can be disregarded.

The author wishes to thank Mr. M. C. Bloom, research associate in geology, for his kind assistance in this investigation.

[7] Stephen S. Visher, "Increased Oceanic Salinity as One Cause of Increased Climatic Contrasts," *Geol. Soc. Amer. Bull. 32* (1921), p. 430.

DISCUSSION: THE DISINTEGRATION AND EXFOLIATION OF GRANITE IN EGYPT

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Houston, Texas

The thesis[1] that unloading of rock masses "is the only agency which appears to have been operative in all cases of exfoliation" (p. 635) is contradicted by the relations of disintegration and exfoliation and of nondisintegration and nonexfoliation of the coarse Syene red granite within historic time in Egypt.[2]

Where surfaces, fresh in the days of the ancient Egyptians, have been exposed to moist or relatively moist conditions, disintegration and exfoliation of the granite have taken place to depths of 5–8 mm. Where comparable surfaces, fresh in the days of the ancient Egyptians, have been exposed only to dry or relatively much drier conditions, no disintegration or exfoliation has taken place. Disintegration and exfoliation are the rule under two types of conditions: (1) where the surface of a granite block has been exposed to the air in Lower Egypt and (2) where the surface of the block has been buried in the soil in Middle and Lower Egypt. The exposed surfaces on the granite blocks forming the facing of the pyramids of Gizeh (Lower Egypt) show exfoliation to a depth of 5–8 mm.; and at the entrance to the temple of the Second Pyramid the degree of disintegration below what would seem to be an old soil line is much greater than above. In the temples of the Luxor district (Middle Egypt) the pedestal, the feet, and lower part of the legs of statues commonly show much exfoliation, although above the knees the original polished surface is still intact, and no incipient disintegration or exfoliation can be detected under tapping with a knife. The same contrast between freshness of the surface above and

[1] Rollin Farmin, "Hypogene Exfoliation in Rock Masses," *Jour. Geol.*, Vol. XLV (1937), pp. 625–35.

[2] Donald C. Barton, "Notes on the Disintegration of Granite in Egypt," *Jour. Geol.*, Vol. XXIV (1916), pp. 382–93.

disintegration and exfoliation below is shown by the granite pillars and walls of the temples. In the dry underground chambers of the Serapeum at Sakkara on the upland the huge sarcophagi of Syene granite still retain their original polish and show no trace of incipient disintegration or exfoliation under tapping. In the Aswan district of Upper Egypt the ancient Egyptians left many half-hewn blocks and discarded fragments in the ancient quarries, from which the granite was obtained for those statues and blocks in Middle and Lower Egypt. Those Egyptians dressed back many "boulders" of exfoliation and carved datable inscriptions in fresh granite. Surfaces in the Aswan district fresh in the days of the ancient Egyptians, are still fresh and show merely an infinitesimally thin film of tarnish and alteration.

The moisture in the air at Aswan is very much less than at Gizeh. At Aswan there is no rainfall, a light dew may fall at night, and the relative humidity at 8:00–9:00 A.M. ranges from 28 to 58 per cent and averages 39 per cent. Gizeh has several light showers a year, has a moderately heavy dewfall, and at 8:00–9:00 A.M. has a relative humidity ranging from 64 to 87 per cent and averaging 72 per cent. The Serapeum at Sakkara lies high above the Nile Valley, in the edge of the plateau, under desert conditions. The chambers, lying deep underground and being sealed from the air, are dusty dry. The temples at Luxor remained partially buried in the damp alluvium of the Nile Valley for more than a millennium. The following observation in regard to connection between soil moisture and disintegration of the granite was made by Daressey of the Service des Antiquités of Egypt:

> Les granites exposés continuellement à l'eau ou au soleil se conservent bien, mais où ils se degradent, c'est lors qu'ils ont enfouis dans un sol humide: la formation de sels, nitrates, et autres, fait alors décomposer le granit, surtout lorsque le terrain est alternativement sec et humide.[3]

In brief, therefore, the disintegration and exfoliation of the granite in Egypt have taken place only if the granite has been exposed to the relatively moist air of Lower Egypt or to moist soils. Granite surfaces which have been exposed only to the drier air of Middle

[3] G. Daressey, personal letter, September 5, 1915.

and Upper Egypt or in the dry air of the Serapeum are still as fresh as they were in the days of ancient Egypt.

Three conclusions follow definitely from these data on the exfoliation in Egypt: first, that this exfoliation in Egypt cannot be even partly hypogene, for if an appreciable part of it were hypogene, traces of disintegration and exfoliation should be recognizable over the surface of those granite sarcophagi, over the fresh surfaces of these blocks in the Aswan district and over the upper part of those statues; second, that the disintegration and exfoliation are directly or indirectly the effect of moisture; and, third, that if nonhypogene exfoliation can take place to a depth of 5–8 mm. in a period of two thousand to five thousand years in Egypt as an effect of moisture, much of the extensive known disintegration and exfoliation of granite in the moist temperate regions of the world is not a result of unloading of the rock but an effect of moisture, presumably hydration plus incipient further alteration.

REVIEWS

Soils: Their Origin, Constitution, and Classification. By G. W. ROBINSON. 2d ed. London: T. Murby & Co., 1936. Pp. 442; figs. 17; pls. 5. 20*s*.

Geologists as a group have been inclined to consider the subject of soils as outside of their immediate realm, and yet the recent interest in soil erosion and more detailed studies of sediments show that actually some knowledge of soil science should be part of every geologist's background of knowledge. An excellent source of information on modern soil science (pedology) is Professor Robinson's book. The first half of the volume is devoted to a review of the factors involved in the soil-forming process, including such topics as weathering, the clay complex, base exchange, soil organic matter, and the water relationships of soils.

The last half of the book is devoted to the main soil groups, including the podsols, the tshernosems, saline soils, tropical soils, etc. This section contains a wealth of informative material that cannot be ignored by the well-informed geologist. At the end of the volume is a discussion of soil classification and world soil distribution. The material is lucidly presented, and each chapter has appended a selected bibliography, which will be welcomed by readers whose interest lies in particular fields of the subject.

W. C. KRUMBEIN

The Story of Yellowstone Geysers. By CLYDE MAX BAUER. St. Paul: Haynes, Inc., 1937. Pp. 125; photographs 99; drawings 3. $1.50. Postage $0.20 extra.

Dr. Bauer, park naturalist of Yellowstone National Park, has utilized his geologic experience and the extensive files of Haynes, Incorporated, authorized photographers of the park, to bring out a very instructive and attractive volume on the Yellowstone geysers. The text is in nontechnical language written in an attempt to answer, simply but authoritatively, many of the questions asked by visitors who wish an explanation of the geysers and their behavior. The illustrations are profuse and beautiful, selected from the best work of two generations of photography in the park.

At the end of the volume is an alphabetical reference list of one hundred and eleven geysers with a few of the most important facts about each. A map of Yellowstone Park locating the different geyser basins, mud volcanoes, hot springs, etc., is included.

R. T. C.